Volume 1, Number 2 April 1993

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

Official English Translation of *Stratigrafiya, Geologicheskaya Korrelyatsiya*

Editor-in-Chief
Boris S. Sokolov

Academician and Councellor of the Presidium
Russian Academy of Sciences

Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

INTERPERIODICA PUBLISHING

Stratigraphy and Geological Correlation

(*Stratigrafiya. Geologicheskaya Korrelyatsiya*)

Published 6 times a year
by Interperiodica (Russia) and the Russian Academy of Sciences
Distributed worldwide by Interperiodica
ISSN: 0869-5938

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 3 - 12.
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Evolution of the biosphere

Dynamics of the Taxonomic Diversity of Riphean and Vendian Stromatolites in Northern Eurasia

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Manuscript received March 18, 1992

Abstract – The sample of Riphean-Vendian stromatolites from Northern Eurasia is very representative for this time interval of geologic history. It includes 262 taxa of species and 70 taxa of genus ranks. An analysis of the systematic compositional changes of these stromatolites at species and genus levels within 7 time units (two each for the Early, Middle, and Late Riphean, and one for the Vendian) has revealed that their diversity grew gradually in the Early and Middle Riphean and sharply increased, particularly at the species level, in the first half of the Late Riphean. A considerable diversity decrease occurred in the second half of the Late Riphean, ca. 850 Ma ago, and continued in a smoothed form in the Vendian. Thus, the Late Proterozoic decline of stromatolites took place considerably earlier than is accepted and could not be related to the reaction of microbial communities to the appearance of Metazoa. The decline coincided chronologically with a number of significant abiotic and biotic events and was one of the expressions of an important rearrangement of Proterozoic ecosystems ca. 850 Ma ago.

INTRODUCTION

Secular variations in the abundance and diversity of stromatolites in the course of geological history attracted geologists' attention for a long time. V.P. Maslov (1959), who was the first to recognize this phenomenon, considered the Proterozoic, Cambrian, and Ordovician to be the time of maximum abundance of stromatolite. A. Fischer (1965) and P. Garrett (1970) wrote about the sharp decrease in the stromatolite content after the Early Ordovician, and I.K. Korolyuk and A.D. Sidorov (1965, 1971) identified the stage of flourishing and maximum taxonomic diversity of fossil stromatolites within the time limits of the Riphean to Early Cambrian. B.M. Keller and M.A. Semikhatov (1968) have referred this stage to the Riphean and Vendian, having noted the impoverishment of stromatolite assemblages in the Vendian and especially in the Cambrian, and, after S.M. Awramik's work (1971), the conclusion about the Vendian stromatolite decline became a truism. Sometimes not only the decrease in abundance and diversity of stromatolites was assumed to be related to this decline, but also the increase in the relative amount of layered (stratiform) and nodular morphs, at the expense of columnar ones, and the change in the dominant microstructure type (Semikhatov, 1974).

Awramik (1971) made the first attempt to estimate quantitatively the secular change in the systematic diversity of stromatolites during the Proterozoic and Cambrian. Considering the time ranges of 70 formal species of columnar morphs, he has demonstrated that their diversity grew from pre-Riphean time to the Late Riphean and then sharply decreased in the Vendian and Early Cambrian. Later, M.R. Walter and G.R. Heys (1985) revised this estimate on a wider database. Using the data on the distribution of 485 taxa of columnar, layered, and nodular morphs, they proposed the now well-known model of dynamics of not only diversity, but also abundance of Proterozoic stromatolites. According to this model, the abundance reached its maximum in the Late Riphean, and the diversity, in the Middle Riphean; it did not change in the Late Riphean and then sharply decreased in the Vendian and Cambrian.

The correctness of the conclusions concerning the secular diversity dynamics in any group of organic remains depends on three factors: the representativeness of the analyzed sample, the reliability and resolution of the stratigraphic framework, and the state of classification of the group.

The sample used by Walter and Heys may be considered as representative. It included the data on Australia, Africa, and North America; the bulk of the data on Northern Eurasia; and a part of the data on China. As for the stratigraphic base of their model, it suffers from two drawbacks. First, the Riphean part of the scale was divided into only three time intervals (Early, Middle, and Late Riphean), and so the analysis could be made with a step of only an era range, and this, as is

shown below, influenced the conclusions about timing of the Late Proterozoic crisis of stromatolites. Second, the Middle-Upper Riphean boundary in Siberian stromatolite-bearing sections, in accordance with the opinions prevailing in the 1960s and 1970s, was placed stratigraphically higher than it should be, and so one of the richest, the so-called Lakhanda stromatolite assemblage of Siberia, was regarded within the Middle rather than Upper Riphean. As a result, the quantitative relationships of the Middle and Late Riphean samples proved to be significantly distorted. The classification base used in the Walter and Heys's contribution was not uniform. A natural attempt to embrace material as comprehensively as possible resulted in including into calculations not only valid species that are more or less adequate in their volumes, but also the indiscerned synonyms and taxa identified by specialists of different schools, which, consequently, are different both in the sets of their diagnostic features and in their volumes. Therefore, noise, hardly yielding to estimation, has been introduced into the statistics.

To minimize the above-mentioned errors in analyzing the dynamics of the diversity of Riphean and Vendian (Late Proterozoic) stromatolites, we have chosen the stromatolites of Northern Eurasia for this analysis. In abundance, wideness of distribution, and the extent to which they have been studied, the stromatolites of Northern Eurasia are first among all the samples of the same age, and the morphs of different types from all parts of the sequence have been studied there in approximately the same details. The paleotectonic and paleoenvironmental settings of the Northern Eurasia Riphean and Vendian deposits were widely variable, and this substantially levels, if not completely eliminates, the influence of regional factors. The stratigraphic base of the Northern Eurasia Upper Proterozoic makes it possible to trace in it not only the Vendian and three Riphean erathems, but also the units subordinate to the latter. It is also significant that the classification of the Riphean and Vendian stromatolites of Northern Eurasia is treated by most of the workers with a sufficient uniformity. Microstructural features serve as a leading or a single criterion for establishing formal species (forms) of the stromatolites, and when identifying formal genera (groups), attention is concentrated on a combination of morphological features of the buildups, sometimes supplemented by the general characteristics of the microstructure (for details, see Krylov, 1975; Raaben, 1986; Semikhatov and Komar, 1989).

In discussing the classification base of the executed analysis, it should be noted that we are well acquainted with the type material of almost all stromatolite taxa established in the Precambrian of Northern Eurasia. This has given us an opportunity to reveal some synonyms, to consider several taxa invalid, and to specify the position of a number of genera in the system of supergeneric categories. All of the above has made it possible to analyze the dynamics of the diversity of the Riphean and Vendian stromatolites from a very representative sample, not only at species, but also at higher taxonomic levels, and with a greater time resolution than was done earlier.

RESULTS OBTAINED

The analyzed sample includes 248 formal species and 70 formal genera of stromatolites from the Riphean, Vendian, and Lower Cambrian of Northern Eurasia, described in the literature and recognized by us as valid. In addition, the sample was replenished at a specific level with 14 taxa, at the expense of such stromatolites that have been either named only in collection or determined in publications only at a formal generic level but having, however, a specific microstructure. Species determined in open nomenclature were not taken into account. The lateral distribution of more than two thirds of the considered formal species (forms) is restricted to the Siberian stromatolite provinces. The data on the vertical distribution of stromatolite taxa was obtained from a number of generalizing publications and monographs (for bibliography, see Dol'nik, 1982; Semikhatov, 1985, 1991; Semikhatov and Komar, 1989), supplemented with the materials of particular papers, and, when necessary, corrected in accordance with modern stratigraphic scales. In addition, the data on the Lower Proterozoic (pre-Riphean) stromatolites of Northern Eurasia has also been considered. The sample consisted of 43 species and 23 genera (Makarikhin and Kononova, 1983; Medvedev, 1991; for other references, see Semikhatov and Komar, 1989). It cannot be regarded as representative for a general analysis of the dynamics of the stromatolite diversity in the relevant time span: throughout Northern Eurasia, the Early Proterozoic stromatolites are distributed sporadically both in time and space, and beyond the bounds of the Yatulian in Karelia, they are still insufficiently studied.

A stratigraphic groundwork of the executed analysis was a more detailed version of the scale of the general subdivision of the Precambrian adopted in 1990 for Northern Eurasia within the limits of the former Soviet Union (Semikhatov *et al.*, 1991). The scale, as is known, envisages a twofold subdivision of the Proterozoic and a subdivision of the Upper Proterozoic (from 1650 ± 50 Ma ago to the base of the Cambrian) into three Riphean erathems (Lower, Middle, and Upper Riphean) and the Vendian System overlying them. The isotopic age of the lower boundaries of the Lower, Middle, and Upper Riphean and the Vendian is assumed to be 1650 ± 50, 1350 ± 20, 1000 ± 50, and 650 ± 20 Ma, respectively (Semikhatov *et al.*, 1991). For the present analysis, the scale was specified and two subordinate units were identified in each of the Riphean erathems. A debatable problem concerning the position of the Middle-Upper Riphean boundary in Siberia (for review and bibliography, see Semikhatov and Serebryakov, 1983; Komar, 1990) was solved in the following manner. According to both the data on the vertical distribution of significant microfossil taxa and some stromatolites, and the available isotopic datings (though methodologically

imperfect), the boundary was drawn at the bases of the Lakhanda Group in the Uchur-Maya hypostratotype of the Riphean and of this group's time analogues (Semikhatov and Serebryakov, 1983; Semikhatov, 1985, 1991). In Southern Siberia, where Riphean deposits have no isotopic datings and contain endemic sets of stromatolites, the boundary, according to the latest data on microfossils (Stanevich *et al.*, 1990), was placed at the base of the Dal'nyaya Taiga Group, though other opinions are also actively argued (e.g., Dol'nik, 1982; Khomentovskii *et al.*, 1985).

The units subordinate to the Riphean erathems, which we used in our analysis, had been already identified in the literature. In Siberia they correspond to the Lower Riphean Chumnakh and Tomulakh horizons of the Anabar massif (Shpunt *et al.*, 1982) and to the Middle and Upper Riphean Aimchan, Kerpyl', Lakhanda, and Ui groups of the Uchur-Maya Riphean hypostratotype. The four mentioned groups are usually considered to be either the types or characteristic representatives of stratigraphic subdivisions common to the entire or almost entire Siberian craton and its surroundings (Khomentovskii, 1976; Semikhatov and Serebryakov, 1983; Khomentovskii *et al.*, 1985). In this paper, we correlated with the Ui and Lakhanda groups the South Siberian Zhuya and Dal'nyaya Taiga groups, respectively, and their counterparts (Stanevich *et al.*, 1990). We did not identify the subordinate units in the Riphean erathems in the pre-Dal'nyaya Taiga deposits of this region, which are very poor in stromatolites. There is no direct evidence on the isotopic age of the boundary between the Chumnakh and Tomulakh horizons, but the K-Ar and Rb-Sr datings of the glauconite group minerals from the lower Chumnakh and post-Tomulakh deposits suggest that its age is close to 1400 - 1450 Ma. As for the boundaries between the subordinate units within the Middle and Upper Riphean erathems, they are usually dated as ca. 1200, 1000, and 850 Ma (Khomentovskii, 1976; Semikhatov and Serebryakov, 1983; Khomentovskii *et al.*, 1985).

In the Uralian stratotype of the Riphean, the counterparts of the considered Siberian units subordinate to the Riphean erathems have been identified with a different degree of confidence. The two units of the Siberian Upper Riphean, according to a complex of paleontological and isotopic geochronological data, correlate well with the Min'yar-Uk and pre-Min'yar parts of the Karatau Group, the Upper and Lower Karatavian. The boundary between the Kerpyl' Group of the Riphean hypostratotype and the underlying Aimchan Group, according to the distribution features of stromatolites and rare isotopic datings, may be provisionally drawn in the middle part of the Avzyan Formation of the Uralian Yurmata Group, and the boundary between the Chumnakh and Tomulakh horizons, according to some distribution features of stromatolites, may be tentatively correlated with the boundary between the Satka and Bakal formations of the Uralian Burzyan Group. At present, we cannot be completely confident that, placed in such a way, the Chumnakh-Tomulakh and Aimchan-Kerpyl' boundaries are isochronous in both Siberia and the Urals. However, the ranges of possible time transgressions of these boundaries, when correlating the Siberian and Uralian sections, seem to be insignificant in comparison with the time ranges of the bounded units.

The resolution of the adopted stratigraphic framework permitted us to notice some features of the dynamics of stromatolite diversity, which escaped the attention of our predecessors, and to equalize the time ranges of the Riphean and Vendian units under comparison.

The results of the analysis can be seen in Fig. 1, where histograms show the number of formal species and genera of stromatolites present in the selected time intervals of the Northern Eurasia sequences, and circle diagrams reflect the percentage of taxa belonging to the usually identified supergeneric categories (Raaben, 1986). The curves of the numbers of appearing and disappearing taxa are shown as well; they are based on the data obtained both within Northern Eurasia and beyond its bounds. Because of this fact and the cases of discrete distribution of taxa in Eurasian sections, the difference between the number of taxa shown in adjacent histogram bars does not always coincide with the difference between the number of taxa appearing and disappearing during the relevant time span.

Figure 1 demonstrates that the secular changes in the diversity of stromatolites of Northern Eurasia in the Late Proterozoic were of the same trend at species and genera levels but were more contrasting at a species level. The total number of species tended to grow gradually from the beginning of the Early Riphean to the end of the Middle Riphean and then, in the first half of the Late Riphean (in the Early Karatavian), they sharply increased. As a result, the total number of species in the Early Karatavian, ca. 1000 - 850 Ma ago, was doubled in comparison with the terminal unit of the Middle Riphean (93 and 45 species, respectively).

In the middle of the Karatavian, ca. 850 Ma ago, the general trend of the secular change in the stromatolite taxonomic diversity was replaced by a reverse one. At this time, the first sharp intensification of the species extinction could be registered in the Proterozoic history of Northern Eurasia. From the Early to Late Karatavian, 88 species did not pass. This extinction was greater than all the earlier ones in the Riphean of Northern Eurasia taken together and was accompanied by an appreciable decrease in the number of species appearing in the Late Karatavian. As a result, the Late Karatavian in Northern Eurasia was a time of a considerable decline in the stromatolite specific diversity, and the Early Karatavian, a time of their maximum diversity. It is significant that the contrasts in the diversity of the stromatolites under consideration in the Middle Riphean and Early and Late Karatavian will remain valid, if correlations of Siberian and Uralian sequences, alternative to those adopted in this paper, are used (e.g., Dol'nik, 1982; Khomentovskii *et al.*, 1985).

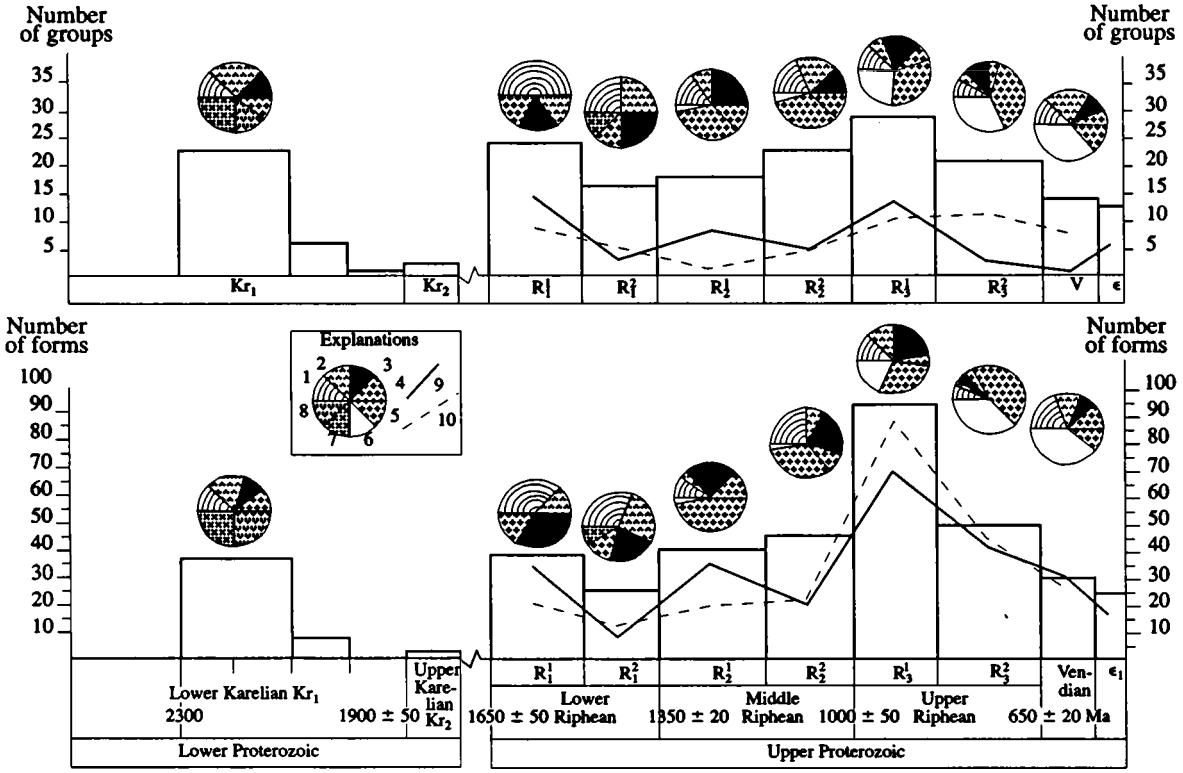

Fig. 1. *Dynamics of the taxonomic diversity of Riphean and Vendian stromatolites in Northern Eurasia at the levels of forms (formal species) and groups (formal genera). 1 - 8 – percentage of the stromatolite species and genera, belonging to supergeneric categories: 1 – layered morphs, 2 – nodular morphs, 3 – conophytonids, 4 – kussiellids, 5 – tungussids, 6 – gymnosolenids, 7 – ministromatites, 8 – intricotids; 9 – number of appearing taxa; 10 – number of disappearing taxa. Histograms show the number of taxa present within indicated time spans of the Precambrian sections of Northern Eurasia.*

In post-Riphean time, the general diversity of stromatolites in Northern Eurasia continued to decrease and in the Vendian, approached the minimum values for the entire Late Proterozoic. The changes in the generic composition of the stromatolites under consideration were, as was already mentioned, identical in direction to those in the specific composition and were inferior to them in amplitudes. Looking at Fig. 1 more closely, it is possible to notice that the number of appearing taxa was appreciably greater in the first half of each Riphean era than in the second half. It is pronounced most strikingly at the specific level, particularly in the Late Riphean.

The considered variations in the diversity of the stromatolite assemblages were accompanied by transformations in their constitution and, in the first place, by relative redistribution of the three main types of buildups: layered, nodular, and columnar (Fig. 2). The percentage of species and genera of the latter was steadily increasing during the Riphean and resulted in their domination of the assemblages by the beginning of the Middle Riphean. The proportion of nodular and layered morphs, rather large in the Early Riphean, decreased to 5 - 10% in the Late Karatavian. Thus, it is possible to talk about the Riphean expansion of columnar forms. However, in the Vendian and Cambrian time, the relative role of columnar stromatolite taxa decreased, while that of layered and nodular ones began to grow again. This growth, as applied to layered forms, was noted earlier (Semikhatov, 1974; Walter and Heys, 1985). In other words, the columnar stromatolites proved to be the most sensitive indicators of ecosystem rearrangements in the Riphean and Vendian.

Along with the secular changes of the columnar stromatolite proportion in the assemblages, the shares of supergroup taxa that were subordinate to them were also redistributed (Fig. 1, circular diagrams, and Fig. 3). Kussiellids and conophytonids were dominant among the columnar stromatolites of Northern Eurasia in the Early Riphean; tungussids with conophytonids became most abundant in the Middle Riphean; in the Early Karatavian, the dominants were equally represented by conophytonids, tungussids, and gymnosolenids, whereas in the Late Karatavian, only tungussids and gymnosolenids remained among the leaders, and in the Vendian and Cambrian, *Boxonia* and related groups assigned to gymnosolenids became an individual leading assemblage. It should be particularly noted that in Northern Eurasia, the ratio of the species number of such characteristic genera as *Conophyton* and *Jakutophyton* to the total number of stromatolite species was also changing during the Riphean and Vendian time. In the beginning of the Early Riphean, it amounted to only 8%; in the Middle Riphean and the Early Karatavian, it increased considerably and varied within narrow limits

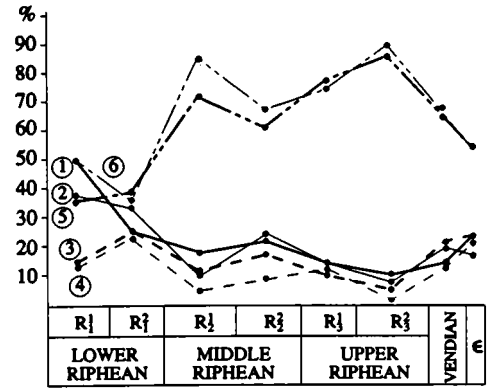

Fig. 2. *Secular variation in the percentage of genera and species of layered, nodular, and columnar stromatolites of the Riphean and Vendian in Northern Eurasia. 1, 2 – genera (1) and species (2) of layered morphs; 3, 4 – genera (3) and species (4) of nodular morphs; 5, 6 – genera (5) and species (6) of columnar morphs.*

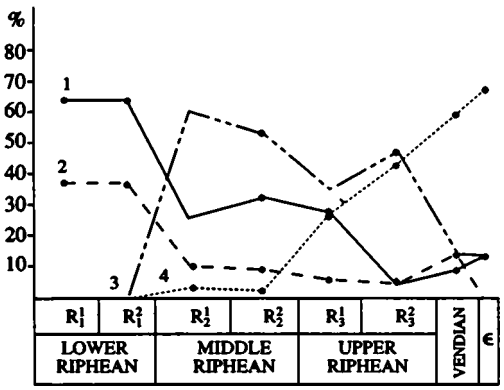

Fig. 3. *Secular variation in the percentage of various supergeneric categories of columnar stromatolites of the Riphean and Vendian in Northern Eurasia. 1 – conophytonids; 2 – kussiellids; 3 – tungussids; 4 – gymnosolenids.*

(20 - 17%), and in the Late Karatavian and in the Vendian, it dropped to 4 - 6%. In the Cambrian of Northern Eurasia, there were no *Conophyton* or *Jakutophyton*.

DISCUSSION

The obtained pattern of the secular variations in the Late Proterozoic stromatolite diversity differs from that published by Walter and Heys (1985) in two main features. First, there is no Middle-Late Riphean plateau in our curve, which some authors considered as very significant (Awramik, 1991; Golubic, 1991); instead, there is a clearly pronounced Early Karatavian peak of diversity. Second, a drastic impoverishment of the stromatolite taxonomic composition, according to our data, began in the middle of the Late Riphean, ca. 850 Ma ago, rather than at the transition from the Late Riphean to the Vendian. This impoverishment was dated by Walter and Heys (1985) as 700 - 800 Ma, and by Walter (1989) in his later publication as 700 Ma, though a more argumented estimation of the isotopic age of the Upper Riphean-Vendian boundary is 650 ± 20 Ma (Semikhatov *et al.*, 1991). It is noteworthy that a substantial drop in the diversity of stromatolites in the middle of the Late Riphean (at the transition to the Sinian) has been also registered in China by Zhu Shixing (1982) and confirmed by our preliminary statistical analysis.

Searching for the possible causes of the above-described secular changes in the diversity of the Riphean and Vendian stromatolites, we emphasize the following:

1. These changes cannot be an artifact related to inadequate information concerning the composition of the succesive stromatolite assemblages or to inadequate understanding of the volumes of species from different horizons of the sequence.

2. Stromatolite taxa, identified according to their microstructures, and, to a degree, those identified according to the combination of morphological peculiarities, reflect mediately the phenotypic and, perhaps, genotypic features of stromatolite-forming microorganisms (Semikhatov *et al.*, 1979; Awramik, 1991; Golubic, 1991). This follows from such well-known facts as: (a) there is a certain correlation between the systematic composition of stromatolite-forming mats, on the one hand, and the microstructure and, partly, some morphological features of stromatolites (e.g., laminae shape), on the other; (b) the structure and function of stromatolite-forming microbial communities are determined by very few species or even by a single species of microorganisms; and (c) each of such species is characterized by its own genetically determined habitat (Serebryakov, 1975; Semikhatov *et al.*, 1979; Awramik, 1991; Golubic, 1991). Therefore, the distribution features of the stromatolite lower taxa reflect to some extent the distribution features of systematically different microbiotas.

3. Within each time interval, the diversity of the stromatolite species identified according to their microstructure did not coincide with the diversity of stromatolite-forming microbial communities that really existed. This was related, in the first place, to the diagenetic and epigenetic processes that modified the original microstructure, either obliterating its features providing a basis for species identification, or, on the contrary, creating some peculiarities corresponding only to the degree of its preservation. At the same time, the build-ups, assigned to different genera because of morphological differences resulting from environmental pressure, may have the same microstructure. These facts, however, cannot appreciably influence the conclusions concerning the dynamics of the stromatolite diversity whatsoever, because the conclusions are based on the analysis of relative, rather than absolute, values.

In looking for the possible causes of the secular changes in the stromatolite diversity, attention was usually concentrated on the explanation of this diversity decline. It was assumed to be related to the either competitive exclusion of stromatolite-forming communi-

ties by benthic algae (eucaryotes) (Fischer, 1965; Monty, 1973; Pratt, 1982), the pressing of these communities into specific, laterally restricted environments by grazing and burrowing animals (Garrett, 1970; Awramik, 1971, 1991; Gebelein, 1976; Walter and Heys, 1985; Walter, 1989), the changes in chemical composition of sea water and, accordingly, in the regime of carbonate accumulation (Fischer, 1965; Monty, 1973; Sokolov and Fedonkin, 1986; Grotzinger, 1990; Sochava and Podkovyrov, 1992), or, finally, the disturbances in the existing cyanobacterial ecosystems during the Early Vendian glacio-eustatic regression, followed by the impossibility of restoring these ecosystems in the course of the postglacial transgression because of the rapidly evolving Metazoa penetration into favorable ecologic niches (Sokolov and Fedonkin, 1986; Fedonkin *et al.*, 1987; Semikhatov and Komar, 1989).

The second of these explanations has proven to be the most popular. It has arisen considerably as a reflection of the fact that the majority of present-day stromatolites is confined to environments extremely unfavorable for animals. However, the absence of unambiguous bioturbations and casts of Metazoa in the pre-Vendian sediments and their wide appearance in the Vendian (e.g., Fedonkin, 1987) permit using this explanation only for the Vendian impoverishment of stromatolite assemblages; otherwise, one must resort to an ambiguous assumption of a long pre-Vendian history of Metazoa, which has not left any tangible traces (Walter and Heys, 1985). A time gap, inconvenient for supporters of this explanation, between the Late Proterozoic decline of stromatolites and the real appearance of Metazoa in the geological record has already attracted attention (Grotzinger, 1990). Our data proves that this gap, at least in Northern Eurasia, amounted to 200 - 250 Ma. This compels us to relate the decline in stromatolite diversity in the middle of the Late Riphean to other causes instead of the Metazoa appearance.

In looking for these causes in the Northern Eurasia record, attention is attracted to the fact that it was in the middle of the Late Riphean, ca. 850 Ma ago, that an extensive regression occurred here (Keller *et al.*, 1974; Khomentovskii, 1976; Semikhatov and Serebryakov, 1983). It resulted in the emergence of considerable areas, activation of landforming processes, and drastic reduction of carbonate accumulation areas and, therefore, of stromatolite abundance. In the Late Karatavian of Northern Eurasia, appreciable amounts of stromatolites survived only in the Baikal-Patom Highland, Ural-Timan, and the Svalbard areas.

In looking for the mentioned causes on a broader scale, the following four circumstances are of paramount importance. First, the Late Karatavian decline of stromatolites coincided in time with the beginning of the African Glacial Era (Chumakov, 1978, 1985, 1987), whose initial stages left after them widespread glacial rock complexes ca. 900 to 740 Ma old and, doubtless, were marked by a considerable decrease in a carbon dioxide content in the atmosphere and carbonate saturation in the sea water. Second, it has been found in a number of regions that the carbonate rocks of the Upper Karatavian (of the second half of the Upper Riphean) differ from older and younger carbonates both in some lithological and mineralogical features and in a distinct carbon isotopic composition of the carbonates, and appear to reflect a special stage in the general evolution of the Precambrian biosphere (Fairchild *et al.*, 1990; Grotzinger, 1990; Knoll and Swett, 1990). Sometimes the entire Late Riphean is included in this stage (Sochava, 1992; Sochava and Podkovyrov, 1992), but it depends apparently on an insufficient resolution of the used stratigraphic framework. Third, the new data on the initial $^{87}Sr/^{86}Sr$ ratios in Riphean carbonates, recently published (Derry *et al.*, 1989; Pokrovskii and Vinogradov, 1991) and obtained by I.M. Gorokhov from our samples, strongly suggests that it is necessary to revise the established pattern of the secular variations in the strontium isotopic composition of sea water (Veizer *et al.*, 1983; Veizer, 1989). This data proves that during the Early Riphean to Early Karatavian, the $^{87}Sr/^{86}Sr$ ratio in sea water was considerably lower than is assumed in the current literature and began intensely growing only in the Late Karatavian, ca. 800 - 850 Ma ago. Therefore, the relative roles of the two most important sources of the materials delivered into the World Ocean, the continental and oceanic crust, changed in the middle of the Late Riphean. The general balance of the continental and oceanic crust (mantle) inputs had changed by this time, and the share of the continental input noticeably increased. Fourth, the statistics of the global data available in the middle of 1988 executed by J.W. Schopf (1991, Fig. 8), has shown that ca. 800 Ma ago, the curve of the systematic diversity of microfossils began to decline. This curve, reflecting the secular variation in the specific diversity of microfossils, has been constructed on the basis of extrapolation between the stratigraphically oldest and youngest occurrences of a species, and thus is most useful for our purposes. The Late Karatavian decline in microfossil diversity, which is clearly seen in the Schopf's curve, appears to be reflected particularly in the upsection decreasing diversity of organic-walled forms recorded in the Upper Riphean sequences of East Siberia and the Urals ("Stratotip...," 1982; Veis, 1988).

Thus, the decline in the diversity of the Late Riphean stromatolites in Northern Eurasia is chronologically and, most likely, in terms of cause and effect, related to a complex of interdependent and independent biotic and abiotic events. In light of data on the influence of major continental glaciations upon the Earth's external spheres (Chumakov, 1985, 1987), it seems plausible to accept that it was glaciations that played an essential, if not a main, role among these events. At any rate, the fact is that in the middle of the Late Riphean, ca. 800 - 850 Ma ago, the Riphean ecosystems were drastically rearranged. The substantial changes in the taxonomic diversity of stromatolites in Northern Eurasia occurring at this time were one of the expressions of this rearrangement.

The further decrease in the Northern Eurasia stromatolite diversity during the Vendian time was forestalled by the Laplandian Glaciation, the most extensive of the Proterozoic glaciations, but continued under the conditions of a vast postglacial transgression, to which widely distributed sediments, including carbonates, were related. These carbonate deposits, widespread throughout huge areas of Siberia and Northeastern Asia and formed in extensive shallow, partly evaporitic basins with small silicoclastic input, seem to have been very favorable for stromatolite formation. Nevertheless, the above deposits are poor both in the number and in the diversity of stromatolites. The most likely explanation for this paradox is an appearance of soft-bodied Metazoa in the Vendian benthic ecosystems. These grazing and burrowing animals prevented benthic stromatolite-building microbial communities, which experienced acute stress in the course of the Laplandian glacioeustatic regression, from gaining anew a dominant position throughout the whole subtidal zone (Sokolov and Fedonkin, 1986; Fedonkin *et al.*, 1987; Semikhatov and Komar, 1989). It is also possible that other factors made their contribution to the Vendian decline of stromatolites. One is a decrease continuing in this time in the global diversity of microorganisms (Schopf, 1991), and another is the further lowering of carbonate content in sea water (Grotzinger, 1990).

In explaining the left ascending part of the histogram of the Upper Proterozoic stromatolite diversity in Northern Eurasia, attention is attracted to the fact that there are three main steps: the Early Riphean, Middle Riphean, and Early Karatavian, which correspond in time to the three main stages of the Riphean transgression, with a maximum in the Early Karatavian, in Lakhanda time (Keller and Semikhatov, 1968; Semikhatov and Serebryakov, 1983). However, the increase of the paleobasin area within the regions under consideration could not be a single, let alone, main cause of the increasing amount of stromatolite taxa. This follows from a lack of coincidence between the growth ranges of the stromatolite diversity and the mentioned transgression, on the one hand, and the diversity growth at the expense of only columnar buildups, on the other. The abiotic factors, which influenced the growth of the stromatolite diversity, might also include a change in the sea water chemistry (for instance, its saturation with carbonates), resulting at the beginning of the Late Riphean in optimum values for the formation of diverse stromatolite-forming mats (Grotzinger, 1990; Sochava and Podkovyrov, 1992). However, the idea explaining the growth of the diversity of Riphean stromatolites by biotic events seems no less attractive. It looks very symptomatic that, according to the Schopf's data (1991), during the Early Riphean to the beginning of the Late Riphean, the global diversity of microfossils was rapidly increasing and peaked at ca. 950 - 800 Ma ago. In other words, the maximum of the Northern Eurasia stromatolite diversity, confined to the beginning of the Late Riphean and most distinctly registered at a specific, i.e., at a microstructural level, almost coincides chronologically with the peak of the global diversity of Proterozoic microfossils. It is appropriate to remember here that in the Middle Riphean and near the base of the Upper Riphean, two new types of clearly biogenic stromatolite microstructures appear, which were not present in older deposits (Komar, 1989; Semikhatov and Komar, 1989). It is quite probably that the phenomenon under discussion, i.e., the sharp increase of stromatolite diversity in the Early Karatavian, was also related to the appearance of both the eucaryotes in stromatolite-forming communities (Awramik, 1991) and new size categories among the microbes existing earlier. This appearance could enhance the adaptive capabilities of mats with regard to trapping a granulometrically wider spectrum of carbonate particles.

Some reduction of the stromatolite diversity in the second subdivision of the Early Riphean, in comparison with the first one, which can be seen in our histograms, can also be explained. In the Urals it is related to the presence of a relatively rich assemblage of Thyssagetacea buildups in the lower of the subdivisions and its absence in the upper ("Stratotip...," 1982). In the Siberian sequences, this reduction reflects a transition from the environments of an open shelf, with diverse stromatolites, to the tidal flat environments with the predominance of unvaried layered and nodular morphs.

Thus, we can conclude that the ascending part of the histogram of the secular variation in the diversity of the Northern Eurasia Riphean and Vendian stromatolites also reflects an interaction of biotic and abiotic factors of various significance, including those of a very wide range of action.

As for the change in the Early Proterozoic stromatolite diversity in Northern Eurasia (Fig. 1, left), it is completely related to the regional features of distribution and, partly, environmental setting of the carbonate rocks, which are, in this part of the sequence, sharply subordinate to the volcaniclastic deposits. Nevertheless, in analyzing the wider data, it can be seen that in a number of regions beyond the Northern Eurasia, the end of the Early Proterozoic (1800 - 1650 Ma ago) was also marked by a decrease in the carbonate rock and stromatolite abundances and diversity. This was possibly caused by the tendency to a generally higher elevation of continents at that time (Semikhatov, 1974).

Outlining the reasons for the redistribution of the relative abundance of the Riphean and Vendian stromatolite supergeneric categories (Figs. 2 and 3) is more difficult. This redistribution could possibly reflect both the change in systematic composition of the stromatolite-forming microbial mats and the broadening of the adaptive capabilities of microbial communities, in particular, their accommodation to a wider range of subtidal environments. If so, the Riphean expansion of columnar morphs, at the expense of layered ones, and the reversal of this trend in the Vendian could most likely be explained. The bulk of columnar stromatolites in the Late Precambrian was confined to the subtidal zone (Serebryakov, 1975), and in the Vendian, they proved to be within the biotope of rapidly evolving

Metazoa, whereas the layered stromatolites, dwellers of intertidal and supratidal environments, developed outside this biotope.

In conclusion, we stress that the dynamics of the diversity of widespread and very representative Riphean and Vendian stromatolites of Northern Eurasia was determined by the combination and interaction of biotic and abiotic factors, at least a part of which were of global scale. The general significance of the conclusions made on the Northern Eurasia material is confirmed not only by the representativeness of the sample used. The case is more interesting. The above histogram of the secular change in the systematic diversity of the Northern Eurasia stromatolites is generally very similar to the histogram that we compiled for the Upper Proterozoic stromatolites from China, the world's second largest region of the wide distribution of these buildups studied by specialists of a single school. A minute analysis of the statistical data on the Chinese stromatolite diversity and a comparison of this data with that from Northern Eurasia and other continents will be given in a separate paper. Here we would like to note that the above data should result in a revision of the presently accepted views concerning the timing of the Proterozoic crisis of stromatolites and in the specification of its cause.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 13 - 25.
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On the Permian Stratigraphic Scheme of Taimyr

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Manuscript received May 12, 1992

Abstract – A detailed description of the sequences of the Belosnezhka and Tsvetochnaya horizons at the Belosnezhka Creek (northwestern shore of the Taimyr Lake) is given. The former is characterized by abundant foraminifers and brachiopods; the latter contains bivalves and imprints of plants. The Belosnezhka Horizon is correlated to the lower substage of the Kazanian Stage, and the Tsvetochnaya Horizon corresponds to the upper substage of the Kazanian Stage of the Permian Chronostratigraphic Scale, both based on analysis of fossils and correlation with synchronous formations of the Verkhoyan'e and Kolyma-Omolon regions.

The first stratigraphic scheme of Permian sediments of the Taimyr region was worked out by L.A. Chaika (in Migai, 1957). According to this scheme, the Lower Permian comprises the Turuza, Byrranga and Sokolinaya formations, while the Upper Permian includes the Baikura and Chernye Yary formations. The upper part of the Tuffolava Formation was referred to the Triassic. The stratotypes of the formations were not indicated, which caused confusion in the determination of the stratigraphic succession of deposits during mapping of the territory and, accordingly, in the exact stratigraphic position of faunal associations. Hence, the same complex could be attributed to different formations. However, this scheme is still used. N.A. Shvedov *et al.* (1961), investigating fossils from these sediments, singled out paleontological horizons under the names of Chaika's formations (Table 1). Later, V.I. Ustritskii and G.E. Chernyak (1963) changed their views on the age of these horizons: the Turuza Horizon was attributed to the interval from the Upper Carboniferous to the Sakmarian Stage, the Byrranga Horizon, to the Artinskian Stage (the Kungurian Stage), and the Sokolinaya Horizon, to the lower half of the Paikhoian Stage (the Upper Permian). The Baikura Horizon was subdivided into two subhorizons. The lower one was referred to the upper part of the Paikhoian Stage, and the upper one, to the Kazanian Stage. The Chernye Yary Horizon was correlated to the Tatarian Stage. Recently, V.I. Ustritskii (1983) again changed his view on the horizon ages. He correlated the Turuza upper subhorizon to the Asselian and Sakmarian, and, the Byrranga Horizon, to the Artinskian Stage. The lower and larger part of the Sokolinaya Horizon was referred to the Kungurian Stage, whereas the upper part was attributed to the lowermost Ufimian. Subhorizons of the Baikura Horizon were regarded as horizons: the Belosnezhka (lower) and the Tsvetochnaya (upper). The stratotype of horizons is located along the Tsvetochnaya River (Fig. 1). The Belosnezhka Horizon is correlated to the larger lower part of the Ufimian Stage, and the Tsvetochnaya Horizon, to the upper part of the Ufimian and the entire Kazanian stages. The Chernye Yary and Zverinyi horizons (the last one was newly singled out by V.I. Ustritskii) were attributed to the Tatarian Stage. Simultaneously, V.I. Ustritskii carried out correlation of the above-mentioned horizons with Permian sediments of the Kolyma-Omolon region. Thus, the Turuza upper subhorizon and the Byrranga Horizon were compared with the Munugudzhak Horizon; the Sokolinaya Horizon was correlated to the Dzhigdala Horizon; the Belosnezhka Horizon, to the Omolon Horizon (without upper part); the Tsvetochnaya Horizon, to the upper part of the Omolon and Gizhiga horizons; and the Chernye Yary and Zverinyi horizons were considered as analogous of the Khivach Horizon. A year later, V.I. Ustritskii (1984) noted that his correlation of Permian horizons of the Taimyr region with the Permian Chronostratigraphic Scale was not quite reliable. Moreover, the Belosnezhka Horizon was conventionally referred to the Ufimian Stage, and the Tsve-tochnaya Horizon was correlated to the Kazanian Stage. V.G. Ganelin (1984) states his own, quite different standpoint about the age of these horizons (Table 1). He correlates the Upper Turuza and Byrranga horizons to the Asselian, and the Sokolinaya Horizon, to the Sakmarian and Artinskian. The Belosnezhka Horizon is referred to the uppermost Artinskian, Kungurian, and Ufimian stages, whereas the Tsvetochnaya Horizon is considered as an analogue of the Upper Ufimian, Kazanian, and Lower Tatarian. The Chernye Yary Horizon and lower part of Zverinyi Horizon are attributed to the Upper Tatarian. Ganelin correlates these horizons with Permian stratotypes of the Kolyma-Omolon and Verkhoyan'e regions, according to the above concept. Recently, D.S. Kashik *et al.* (1990) attributed, using paleomagnetic data, the Khivach and Gizhiga horizons to the Tatarian Stage; the upper part of the Omolon Horizon, to the Kazanian Stage; and the lower part of the Omolon Horizon and upper part of the Dzhigdala Horizon, to the Ufimian Stage (Table 1).

R.V. Solomina studied the stratigraphy of Permian sediments and brachiopods of the Verkhoyan'e region

for a long time ("Resheniya...," 1982; Solomina, 1989; Shul'gina and Solomina, 1990). These works permitted creating a regional scheme of the Permian. Recently, she investigated the Kolyma-Omolon and Taimyr regions. As a result, correlation of the deposits of these regions with the Permian Chronostratigraphic Standard was carried out (Table 1). It is clear that the horizon age definitions by two different methods, i.e., paleomagnetic (Kashik *et al.*, 1990) and paleontological ("Resheniya...," 1982; Solomina, 1989; Shul'gina and Solomina, 1990), do not agree and require further investigation. The greatest divergences are related to the age definitions of the Belosnezhka and Tsvetochnaya horizons.

The elucidation of the age of the Upper Permian brachiopods (the Belosnezhka Horizon) required additional research, such as a detailed description of sections and sampling of fossils. This was done by Solomina in a sequence along the Nyun'-Karakutari River and Belosnezhka Creek in 1986 and 1987. The Belosnezhka sequence had been described in detail, followed by bed to bed sampling of macro- and microfauna by E.N. Preobrazhenskaya in 1960, and Solomina corrected this description in 1987. Additionally collected brachiopods and bivalves were classified by R.V. Solomina and M.M. Astaf'eva, respectively; G.P. Sosipatrova studied foraminifers, and plant imprints were studied by N.A. Shvedov and M.V. Durante.

A detailed description made by Preobrazhenskaya is cited below for the Belosnezhka and Tsvetochnaya horizons along Belosnezhka Creek (Fig. 1), where the section is most complete. The boundaries relative to the underlying Sokolinaya and overlying Chernye Yary horizons are conformable.

The section consists of thick, terrigenous deposits characterized by clear rhythms. Layers of clayey sandstones or clayey aleurolites with suspending marks, containing the most abundant and diverse marine fauna, are located at the base of the rhythms. Clayey layers are characterized by fine, lens-like sinuous bedding; sandy layers have cross, lens-like and discontinuous bedding. Flatted pebbles of clayey rock and frequent ripple marks are observed on bed surfaces. The sediments are bioturbated throughout the section. The section is intruded repeatedly by basic sills and dikes. Sedimentary rocks are compacted and metamorphosed to hornfels at the contacts. The sequence below is described from the base upward.

THE UFIMIAN STAGE

The Sokolinaya Horizon (upper part). Outcrop 30 is situated at the upper reaches of Belosnezhka Creek, on the left bank of the right tributary, 6 km from the creek mouth (Fig. 2).

1. Gray and light-gray fine-grained and medium-grained sandstones with coaly interlayers. Thickness is 18 m.

2. Thin interlayering of gray and dark-gray argillites, clayey aleurolites, clayey sandstones with pyritic and carbonate concretions, plant detritus (7 m). These sediments are overlain by 33 m-thick fine-grained and medium-grained gray sandstones, which are analogous to those from bed 1, and contain rare interlayers of clayey rocks, sideritic concretions, and plant detritus. Thickness is 40 m.

3. Covered, 5-m-wide interval. The next 20 m shows interlayering gray and dark-gray argillites, clayey aleurolites, and clayey sandstones with pyritic inclusions. The overlying 62 m is composed mostly of light-gray to pinkish-gray fine-grained and medium-grained sandstones with lenses of snow-white varieties. Sandstones contain pyritic and carbonate concretions, massive limy lenses, and plant detritus. A dolerite sill (15 m thick) is observed 20 m below the top. The foraminifers *Frondicularia* aff. *planilata* Gerke and *Nodosaria incelebrata* Gerke and accumulations of the bivalves *Myonia (Myonia)* ex gr. *komiensis* (Masl.), *M. (M.) gibbosa* (Masl.), *Trabeculatia quadrata* (Lutk. et. Lob.), and *Kolymia* sp. are found in the upper part. Thickness is 82 m.

The total thickness of the exposed area of the Sokolinaya Horizon is 140 m.

KAZANIAN STAGE

Lower Substage

Baikurian Superhorizon, Belosnezhka Horizon. Outcrop 31 is situated on the right bank of the creek, 700 m south of outcrop 30.

An evident change in lithology takes place at the Sokolinaya and Belosnezhka horizon boundary.

4. A basal 30-m-thick member of the horizon consists of gray and dark-gray poorly-sorted inequigranular clayey sandstones with brown spots and dispersed pebbles of quartz, jaspers, metasandstones and metaaleurolites, and acid and medium effusives (rarely). Marks of sediment stirring, a presence of limy and phosphate-siderite concretions, and inclusions of pyrite are typical. Rare, indefinable fragments of bivalves and plant detritus are met in the lower part, and remnants of the brachiopods *Orbiculoidea jangarensis* Ustr., *Terrakea* cf. *belokhini* Gan., *Rhynchopora lobjaensis* Tolm., *Olgerdia zavodovskyi* Grig., *Kungaella* cf. *kolymaensis* (Tolm.), and *Brachythyrina sibirica* Tschernjak and the bivalves *Kolymia* cf. *inoceramiformis* Lich. and *?Intomodesma* sp. are present in the upper part. The next 45 m up is composed of dark gray clayey aleurolites with phosphate-sideritic concretions and pyrite intercalations. The brachiopods *Neochonetes* sp., *Strophalosia* cf. *tolli* (Fred.), *Terrakea* cf. *belokhini* Gan., *T.* sp., *Rhynchopora lobjaensis* Tolm., and *Dielasma* sp. and the bivalves *Wilkingia* sp. are found in these sediments. The next 16 m is represented by black argillites with rare interlayers of gray aleurolites having plant detritus and remnants of gastropods. The uppermost 2 m is composed of gray, fine-grained sandstones with numerous charred and ferruginous remnants of

Table 1. The correlation scheme of Upper Permian formations of Taimyr, Kolyma-Omolon and Verkhoyan'e regions

Chronostratigraphic scale			Shvedov, N.A. *et al.* (1961)		Ustritskii, V.I. and Chernyak, G.E. (1963)		Ustritskii, V.I. (1983)			Ustritskii, V.I. (1984)
Period	Epoch	Age	Subepoch	Horizon	Horizon	Subhorizon	Superhorizon	Horizons of Taimyr region	Horizons of Kolyma-Omolon region	Horizons
Triassic				Tuffolava Formation	Lava Formation					Fad'ukud
Permian	upper	Tatarian	upper	the beds with pelecypods and flora imprints	Chernye Yary		Zverinyi Chernye Yary		Khivach	Zverinyi Chernye Yary
		Kazanian		Chernye Yary	Baikura	upper	Baikura	Tsvetochnaya	Gizhiga	Tsvetochnaya
		Ufimian	lower	Baikura		lower		Belosnezhka	Omolon	Belosnezhka
					Sokolinaya					
	lower	Kungurian	upper	Sokolinaya	Byrranga		Sokolinaya		Dzhigdala	Sokolinaya
		Artinskian		Byrranga			Byrranga		Munugudzhak	Byrranga
		Sakmarian	lower	Turuza	Turuza		Turuza upper subhorizon			Turuza upper subhorizon
		Asselian								
Carboniferous										

Table 1. (Contd.)

Ganelin, V.G. (1984)			Kashik D.S. and Ganelin V.G. (1990)	Solomina, R.V.				
				(1990)		(1982)		(1989)
Taimyr	Kolyma-Omolon region	Verkhoyan'e region	Omolon massif	Taimyr	Kolyma-Omolon region	Verkhoyan'e region		
Horizons		Formations		Horizons				
				Zverinyi				
Zverinyi (lower part)	Gagarii		Khivach	Chernye Yary	Gagarii Island			
Chernye Yary			Gizhiga					
	Khivach							
Tsvetochnaya	Gizhiga	Dulgalakh	upper part of Omolon	Tsvetochnaya	Khivach	Dulgalakh	upper	Khal'-pirki
					Gizhiga		lower	Amkan-dzha
				Belosnezhka	Omolon	Delenzha	upper	Nuneg
	Omolon	Delenzha	Omolon	Sokolinaya			lower	Mugochan
Belosnezhka			upper part of Dzhigdala				subhorizons	
	Dzhigdala	Tumara	Dzhigdala		Dzhigdala	Tumara		
		Khabakh	Munugudzhak	Byrranga	Munugud-zhak	Bytantai		
Sokolinaya		Echii						
Byrranga	Munugud-zhak	Kygyltas (upper part)		Turuza upper subhorizon				
Turuza upper subhorizon								

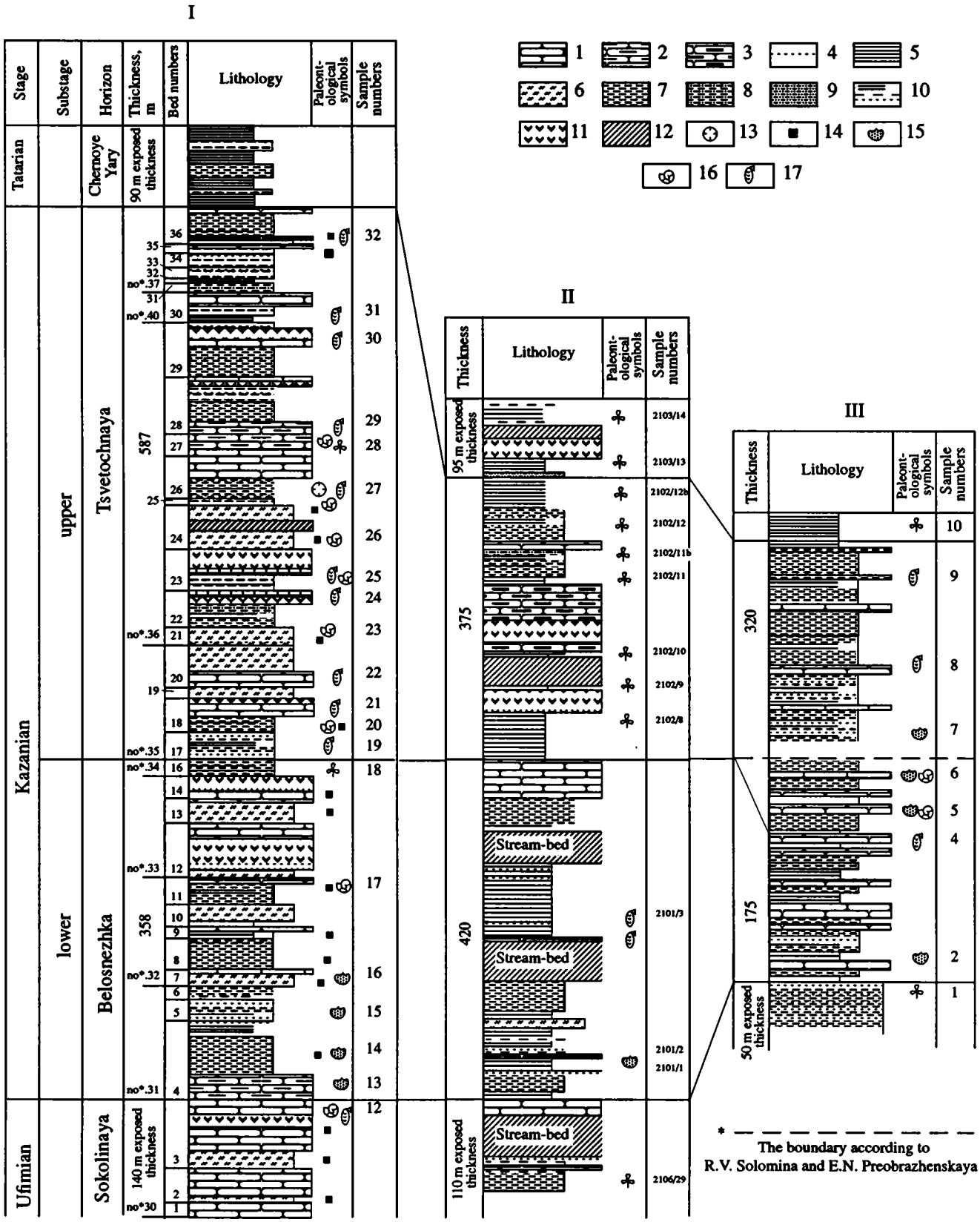
I
II
III
Stage
Substage
Horizon
Thickness, m
Bed numbers
Lithology
Paleontological symbols
Sample numbers
Tatarian
Kazanian
Ufimian
upper
lower
Chernoye Yary
Tsvetochnaya
Belosnezhka
Sokolinaya
90 m exposed thickness
587
358
140 m exposed thickness
Thickness
95 m exposed thickness
375
420
110 m exposed thickness
320
175
50 m exposed thickness
Stream-bed
1 2 3 4 5
6 7 8 9 10
11 12 13 14 15
16 17
The boundary according to R.V. Solomina and E.N. Preobrazhenskaya

Fig. 1. *The scheme of correlation of the Upper Permian sections on the Belosnezhka, Nun'-Karakutari and Tsvetochnaya rivers. Scale 1: 5000. 1 – sandstones; 2 – clayey sandstones; 3 – horizontal bedding; 4 – sandstones of less than 1 m thick; 5 – argillites; 6 – cm-interlayering; 7 – aleurolites; 8 – sandy aleurolites; 9 – bedded aleurolites; 10 – talus; 11 – dolerite sills; 12 – not exposed; 13 – conglomerates; 14 – pyritic crystals; 15 – brachiopods; 16 – foraminifers; 17 – bivalves; no* – outcrop numbers.*

Fauna, I – Belosnezhka Creek (Preobrazhenskaya and Solomina, 1987).

Sample 12 – Frondicularia aff. *planilata* Gerke, *F. mica* Gerke, *Nodosaria incelebrata* Gerke; *Myonia (Myonia)* ex gr. *komiensis* (Masl.), *M. (M.) gibbosa* (Masl.), *Traberculatia quadrata* (Lutk. et Lob.), *Kolymia* sp.; *13 – Orbiculoidea jangarensis* Ustr., *Terrakea* cf. *belokhini* Gan., *Rhynchopora lobjaensis* Tolm., *Olgerdia zavodowskyi* Grig., *Kungaella* cf. *kolymaensis* (Tolm.), *Brachythyrina sibirica* Tschernjak; *14 – Neochonetes* sp., *Strophalosia tolli* (Fred.), *Terrakea* cf. *belokhini* Gan., *Rhynchopora lobjaensis* Tolm.; *Kolymia* cf. *inoceramiformis* Lich.,*? Intomodesma* sp.; *15 – Rhynchopora lobjaensis* Tolm.; *16 – Olgerdia zavodowskyi* Grig., *Cancrinelloides yuregensis* Sol., *Brachythyrina sibirica* Tschernjak, *Rhynchopora lobjaensis* Tolm., *Bajtugania boguchanica* Sol., *Glomospira* ex gr. *gordialis* P. et J.; *17 – Nodosaria krotovi* Tscherd., *Pseudonodosaria* cf. *ventrosa* Schleifer; *18 – Parashisoneura* aff. *sibirica* (Neub.) Radcz.; *19 – Myonia longa* (Lutk. et Lob.), *M. komiensis* (Masl.), *M. bulkurensis* Ast.-Urb., *Schizodus* sp.; *20 – Astartella permocarbonica* (Tschern). *Megadesmus omolonicus* (Mur.) ; *Nodosaria krotovi* Tscherd., *N. incelebrata* Gerke, *N. noinskii* Tscherd., *N. pseudoconcinna* K.M.-Maclay, *Rectoglandulina pygmaeformis* A.M.-Maclay, *Lingulinella* aff. *familaris* Tscherd., *Frondicularia bajcurica* Sossip., *F. mica var. religua* Gerke, *F. planilata* Gerke, *F. pseudotriangularis* Gerke, *F. spectata* Sossip.; *21 – Streblopteria levis* (Lutk. et Lob.), *Polidevcia* sp., *Myonia* sp.; *22 – Streblopteria levis* (Lutk. et Lob.); *23 – Glomospira* ex gr. *gordialis* P. et J., *Dentalina kalinkoi* Gerke, *Nodosaria* sp., *Rectoglandulina* sp.; *Myonia longa* (Lutk. et Lob.), *M. alta* (Lutk. et Lob.); *Pseudonodosaria ventrosa* Schleifer, *Frondicularia* aff. *planilata* Gerke, *Rectoglandulina pygmaeformis* A.M.-Maclay; *24 – Hyperammina* sp., *Frondicularia* sp, *Rectoglandulina borealis* Gerke; *Sanguinolites bicarinatus* Keys., *Streblochondria englechardti* (Eth. et Dun.); *25 – Polidevcia cumboides* (Lutk. et Lob.); *Nodosaria cuspidatula* Gerke, *N. krotovi* Tscherd., *N.* ex gr. *noinskii* Tscherd., *Lingulinella arctica* Gerke, *Frondicularia mica* Gerke, *Rectoglandulina pygmaeformis* A.M.-Maclay; *26 – Myonia longa* (Lutk. et Lob.); *Dentalina kalinkoi* Gerke, *Rectoglandulina pygmaeformis* A.M.-Maclay, *Frondicularia* sp., *Nodosaria incelebrata* Gerke, *N. krotovi* Tscherd. *Glomospira* ex gr. *gordialis* P. et J.; *27 – Streblopteria levis* (Lutk. et Lob.), *Streblochondria englechardti* (Eth. et Dun.), *Solemia biarmica simmetrica* (Lutk. et Lob.) *Myonia longa* (Lutk. et Lob.); *28 – Cordaites* ex gr. *candalepensis* (Zal.) S. Meyen, *Psygmophyllum* sp.; *Nodosaria cuspidatula* Gerke, *Frondicularia spectata* Sossip., *Rectoglandulina* sp.; *29 – Streblopteria levis* (Lutk. et Lob.) *Myonia (Myonia) gibbosa* (Masl.), *M. (M.)* ex gr. *komiensis* (Masl.), *M. (M.) alta* (Lutk. et Lob.), *Streblochondria englechardti* (Eth. et Dun.), *S.* ex gr. *krasnoufimskiensis* (Fred.), *Sanguinolites* sp.; *Beecheria bajkurica* Tschernjak; *30 – Modiolopsis* sp., *Streblopteria* sp., *S. levis* (Lutk. et Lob.), *Polidevcia* sp. *31 – Myonia (Pachymyonia) bicarinata* Ast.-Urb., *Myonia (M.) komiensis* (Masl.), *Streblopteria levis* (Lutk. et Lob.), *Sanguinolites lunulatus* (Keys.), *Polidevcia* sp.; *32 – Streblopteria levis* (Lutk. et Lob.), *S. rotunda* (Lutk. et Lob.), *Streblochondria* cf. *sericea* (Vern.); *33 – Sanguinolites* cf. *bicarinatus* Keys., *Streblopteria levis* (Lutk. et Lob.).

Fauna, II – The Nun'-Karakutari River (Solomina, 1986).

Sample 2106/29 – Koretrophyllites sp., *Paracalamites crassus* Gorel., *P. angustus* Such., *Cordaites* aff. *latifolius* (Neub.); *2101/1 – Strophalosia tolli* (Fred.), *Cancrinelloides yuregensis* Sol., *Brachythyrina sibirica* Tschernjak, *Rhynchopora lobjaensis* Tolm., *Bajtugania boguchanica* Sol.; *2101/2 – Strophalosia tolli* (Fred.), *Cancrinelloides yuregensis* Sol., *Olgerdia zavodowskyi* Grig., *Brachythyrina tricostata* Sol.; *2101/3 – Kolymya* sp., *Allorisma* sp.; *2102/8 – Cordaites* aff. *gracilentus* (Gorel.) S. Meyen, *Glottophyllum* sp., *Sylvella (?)* sp.; *2102/9 – Cordaites* aff. *singularis* (Neub.), *C.* ex gr. *candalepensis* Zal.; *2102/10 – C.* aff. *singularis* (Neub.) S. Meyen, *C.* cf. *candalepensis* (Zal.); *2102/11 – Pecopteris* ex gr. *anthriscifolia* (Goepp.) Zal., *Cordaites* aff. *latifolius* (Neub.) S. Meyen, *C.* aff. *gracilentus* (Gorel.) S. Meyen, *C. candalepensis* (Zal.) S. Meyen, *C.* ex gr. *minax* (Gorel.) S. Meyen. *Lepeophyllum actaeonelloides* (Gein.) Zal., *Samaropsis borisovaensis* Such.; *2102/12 – Crassinerwia kuznetskiana* Neub., *Cordaites* aff. *gracilentus* (Gorel.) S. Meyen, *Lepeophyllum* sp.; *2102/12b – Parashizoneura* sp., *Cordaites* ex gr. *candalepensis* (Zal.); *2103/13 – Phylloteca* cf. *turnaensis* Gorel., *Annularia* cf. *lanceolata* Radcz., *Pecopteris* aff. *anthriscifolia* (Goepp.) Zal., *Cordaites gracilentus* (Gorel.) S. Meyen, *C. chachlovii* (Gorel.), *C. cefaloidea* Mensh.; *2103/14 – Cordaites clercii* (Zal.), *C. angustifolius* (Neub.) S. Meyen, *Lepeophyllum* aff. *actaeonelloides* (Gein.) Zal.

Fauna, III – The Tsvetochnaya River (Ustritskii and Sosipatrova, 1959).

Sample. 1. Noeggerathiopsis taimyrica Schwed.; *2 – Megousia yakutica* (Lich.), *Rhynchopora lobjaensis* Tolm., *Licharewia* sp.; *5 – Orbiculoidea kolymaensis* Lich., *Cancrinelloides ochotica* (Zaw.), *Stepanoviella curvata* (Tolm.), *Strophalosia bajkurica* Ustr, *Rhynchopora lobjansis* Tolm. *Spirifella lita* Fred., *Brachythyrina (?) sibirica* Tschernjak, *Cleiothyridina* cf. *pectinifera* (Sow.); *Nodosaria gigantea* Sossip., *N. krotovi* Tscherd, *N.* cf. *solidissima* Gerke, *Pseudonosaria ventrosa* Schleifer, *Dentalina kalinkoi* Gerke, *Frondicularia inflata* Gerke, *Rectoglandulina pygmaeformis* A.M.-Maclay; *6 – Strophalosia bajkurika* Ustr., *Cancrinella ochotica* (Zav.), *Rhynchopora lobjaensis* Tolm.; *Allorisma* sp., *Solemya* sp.; *Orthovertella* sp. *Nodosaria krotovi* Tscherd., *N. lata* Sossip., *Frondikularia* aff. *mica* Gerke, *F.* cf. *pseudotriangularis* Gerke; *7 – Beecheria bajkurica* Tschernjak; *9 – Tracia alta* Lutk. et Lob.; *10 – Pseudamusium levis* Lutk. et Lob.; *11 – Noeggerathiosis* sp.

plant trunks (up to 0.1 m in diameter). The total thickness of the member is 93 m.

5. The basal part of the member is represented by dark-gray clayey sandstones, black argillites, and gray aleurolites with siderite concretions. Fine-grained gray sandstones (3 m) with lenses of limy rocks, containing the brachiopods *Rhynchopora lobjaensis* Tolm., crinoids, and imprints of plant trunks compose the upper part of the member. Thickness is 24 m.

6. Dark-gray clayey aleurolites and argillites (5 m) with pyrite inclusions. The upper 11 m is represented by interleaved argillites and aleurolites. The quantity of the latter increases upwards, and gray fine-bedded aleurolites (4 m) with accumulations of bivalves and gastropods are present in the uppermost part of the member. Thickness is 18 m.

Outcrop 32 is located on the right bank of the stream, 1 km south of outcrop 31.

7. The basal layer is composed of poorly-sorted clayey sandstones (1.5 m) with the numerous brachiopods *Cancrinelloides yuregensis* Sol., *Olgerdia zavodowskyi* Grig., *Brachythyrina sibirica* Tschernjak, *Bajtugania boguchanica* Sol., and *Rhynchopora lobjaensis* Tolm., bivalves *Allorisma* sp., and foraminifer *Glomospira* ex gr. *gordialis* Parker et Jones. Upward, clayey sandstones are replaced gradually by clayey aleurolites (10 m), with thin layers of argillites and rare homogeneous aleurolites with plant detritus. The overlying 5 m comprises interlayering aleurolites and clayey sandstones. The top of the member is represented by gray, fine-grained sandstones (5 m) with pyrite and rare poorly preserved bivalves and gastropods. Thickness is 20 m.

8. Dark-gray clayey poorly bedded aleurolites (5 m) with shell-like jointing, higher replaced by a 10-m bed of interlayered clayey aleurolites, aleurolites, and argillites. Marks of erosion, clayey pebbles, pyrite inclusions, and plant detritus are observed. The uppermost 4 m is composed of light-gray and brown-gray sandy cross-sinuous coarsely lenticular aleurolites with inclusions of ocherous pyrite, coal-clayey lenses, and limy aleurolites with rare antraconites. Thickness is 40 m.

9. Clayey rocks (5 m) with large lenses of limy aleurolites, clayey pebbles, plant imprints, and pyrite inclusions; rare beds (0.10 - 0.15 m) of black coaly argillites. The upper half (7 m) is composed of gray fine-grained sandstones. Thickness is 12 m.

10. Dark-gray (almost black) argillites (7 m) with pyrite inclusions and shell-like jointing. Further up is a bed (23 m) of aleurolites and sandstones with thin argillite and aleurolite interlayers. Thickness is 30 m.

11. Dark-gray clayey sandstones (0.8 m) with stirring structures gradually changing into clayey aleurolites (6 m) with pyrite inclusions and limy concretions bearing the bivalves *Sanguinolites* sp. and the foraminifers *Glomospira* sp., *Frondicularia* sp., *Nodosaria krotovi* Tscherd., and *Pseudonodosaria* cf. *ventrosa* Schleifer. The middle part is represented by clayey dark-gray aleurolites and argillites with pyrite. Interlayers of gray homogeneous aleurolites appear in the upper part. The large aleurolite lens (2.5 m) containing fragments of the mineralized wood is observed here. The top of the member is composed of gray, fine-grained sandstones (7 m) with numerous clayey pebbles. Thickness is 30 m.

Outcrop 33 is located on the left bank of Belosnezhka Creek, 120 m south of outcrop 32, directly above member 11.

12. The lower 10 meters is represented by interlayered clayey aleurolites and sandstones. There are plant remnants in the aleurolites and coal clasts in the talus. Higher rocks are overlain by light-gray and dark-gray (sometimes pinkish) fine-grained and medium-grained sandstones (16 m) with limy concretions, clayey pebbles, and dispersed pyrite. Rare imprints of plant leaves and trunks, as well as lenses of coal, are present at the top. The bottom of the member is intruded by a dolerite sill (40 m). The latter is underlain by sandstones (2 m)

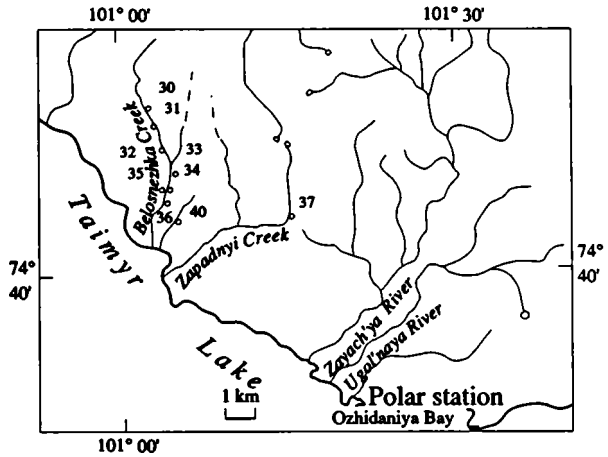

Fig. 2. *The location of the outcrops of the Belosnezhka Creek section.*

and covered by the talus (3 m) in the upper part. The total thickness of the sandstones is 31 m.

13. The member includes interleaved black argillites and dark-gray aleurolites with abundant coalificated plant detritus, siderite concretions, dispersed pyrite, coal lenses (up to 0.3 m), and gray, fine-grained sandstones. The rocks are crowned with sandstone (7 m) containing plant imprints of *Noeggerathiopsis* sp. and *Samaropsis* sp. Thickness is 34 m.

14. Coaly-clayey rocks (1 m) are at the base, light-gray fine-grained sandstones (5 m) with pyrite and abundant clayey pebbles and plant imprints are above. The uppermost part is intruded by the dolerite sill (20 m). Thickness is 6 m.

Outcrop 34 is located on the left bank of Belosnezhka Creek, 500 m south of outcrop 33.

15. Clayey aleurolites (1 m) containing plant imprints of *Parashisoneura* aff. *sibirica* (Neub.) Radcz. Dark-gray cross-bedded aleurolites (4 m) with ripple marks and numerous clayey pebbles. Thickness is 5 m.

16. Separate exposures of clayey aleurolites and argillites (14 m) with coaly rocks, interleaved with homogeneous aleurolites and sandstones. The latter increase upwards in abundance. Lenses of coal (up to 0.1 m) are present. Pinkish-gray sandstones (1 m) are at the top of the member. Thickness is 15 m.

The total thickness of the Belosnezhka Horizon is 358 m.

Upper Substage

Tsvetochnaya Horizon. Outcrop 35 is located on the right bank of the stream, 500 m south of outcrop 34. The next sequence is observed above bed 16:

17. Dark-gray clayey sandstones (2 m) with stirring structure and broken shells. Clayey sandstones change gradually upwards into clayey aleurolites of the same structure, with abundant bivalves *Myonia longa* (Lutk. et Lob.), *M. komiensis* (Masl.), *M. bulkurensis* Ast.-Urb., and *Schisodus* sp. and coalificated and ferruginous plant remnants. The beds of homogeneous aleurolites

and argillites with shell-like jointing and pyrite inclusions appear upwards. Gray fine-grained sandstone beds (0.10 - 0.80 m) with lenses of limy rocks are present in the uppermost 13 m. Thickness is 30 m.

18. Clayey rocks (10 m) similar to those of member 17. There are numerous bivalves *Astartella permocarbonica* (Tschern.) and *Megadesmus omolonicus* (Mur.) and foraminifers *Nodosaria krotovi* Tscherd., *N. incelebrata* Gerke, *N. noinskii* Tscherd., *N. pseudoconcinna* K.M.-Maclay, *Rectoglandulina pygmaeformis* A.M.-Maclay, *R.* cf. *borealis* Gerke, *Lingulinella* aff. *familaris* Tscherd., *Frondicularia (F.) bajcurica* Sossip., *F. (F.) mica* var. *religua* Gerke, *F. (F.) planilata* Gerke, *F. (F.) pseudotriangularis* Gerke, and *F. (F.) spectata* Sossip. The overlying 10 m of aleurolites and argillites contains beds of fine-grained sandstones and pyrite inclusions. The uppermost 14 m consists of gray fine-grained sandstones with the lenses of limy sandstones, pyrite and siderite concretions, and the bivalves *Streblopteria levis* (Lutk. et Lob.), *Polidevcia* sp., and *Myonia* sp. Dolerite sill (8 m) is observed at the top of the member. Thickness is 34 m.

19. Interlayering (0.20 - 0.30 m) clayey aleurolites, aleurolites, and argillites with gray fine-grained sandstones containing limy lenses, with indeterminable gastropods and small bivalves. Thickness is 15 m.

20. Gray and dark-gray clayey sandstones (20 m), gradually changing into clayey aleurolites with dispersed pebbles of various rocks (clayey, sandy, igneous) and abundant bivalves *Streblopteria levis* (Lutk. et Lob.) and *S.* sp. The rocks contain plant detritus and imprints of trunks. The uppermost 39 m is composed of interlayering dark-gray clayey aleurolites, gray aleurolites, and fine-grained sandstones with the bivalve *Myonia sulkatiformis* Ast.-Urb. Thickness is 52 m.

Outcrop 36 is located on the left bank of the stream, 500 m southward from outcrop 35.

21. Clayey sandstones and aleurolites (4 m) with rare beds of gray aleurolites enriched with pyrite are characteristic of the base of the member, containing undeterminable bivalves and the foraminifers *Glomospira* ex gr. *gordialis* Parker et Jones, *Dentalina* aff. *kalinkoi* Gerke, *Nodosaria* sp., and *Rectoglandulina* sp. The upper part (18 m) is composed of a lens-like coarse interlayering of clayey aleurolites (containing pyrite and syderite concretions), argillites, and fine-grained sandstones. The sandstones include fine pebbles and limy concretions with the foraminifers *Pseudonodosaria* cf. *ventrosa* Schleifer, *Frondicularia (F.)* aff. *planilata* Gerke, and *Rectoglandulina pygmaeformis* A.M.-Maclay and bivalve *Myonia longa* (Lutk. et Lob.), as well as imprints of starfish and coalificated and ferruginous plant remnants. Thickness is 22 m.

22. Dark-gray clayey aleurolites (3 m) with the foraminifers *Hyperammina* sp., *Frondicularia (F.)* sp., and *?Rectoglandulina borealis* Gerke, lenses with gastropods, broken bivalves, and rare imprints of cordaites are present at the base. The upper part (24 m) is represented by clayey aleurolites and argillites with beds of homogeneous aleurolites. The top of the member (7 m) consists of aleurolites grading into fine-grained sandstones with two beds of black coaly argillites and lenses of coal. The bivalves *Sanguinolites bicarinatus* Keys. and *Streblochondria englechardti* (Eth. et Dun.) are found here. The uppermost part is intruded by sill (9 m) with globular jointing. Thickness is 34 m.

23. The composition of this member is similar to that of member 22. The middle part contains the foraminifers *Nodosaria cuspidatula* Gerke, *N. krotovi* Tscherd., *N.* ex gr. *noinskii* Tscherd., *N. panda* Sossip., *Lingulinella* cf. *arctica* Gerke, *Frondicularia (F.) mica* Gerke, and *Rectoglandulina pygmaeformis* A.M.-Maclay and the bivalve *Polidevcia cumboides* (Lutk. et Lob.). The upper part is intruded by dolerite sill (25 m). The thickness of sedimentary rocks is 24 m.

24. Clayey aleurolites interbedded with siltstones, sandstones and argillites (rarely). The quantity and thickness of the sandstone and aleurolite beds increase upward. Limy lenses, dispersed pyrite, clayey and siliceous pebbles, gastropod remnants, and the bivalves *Myonia* sp. are found. Ten meters is not exposed. The next 20 m is represented by a clayey bed with aleurolite and sandstone layers, which ends with a bed of fine-grained sandstones (1 m) containing antraconites. The lower part contains the bivalve *Myonia longa* (Lutk. et Lob.). The upper part contains the foraminifers *Dentalina* aff. *kalinkoi* Gerke, *Rectoglandulina pygmaeformis* A.M.-Maclay, *R.* cf. *borealis* Gerke, *Frondicularia (F.)* sp., *Nodosaria incelebrata* Gerke, *N. krotovi* Tscherd., and *Glomospira* ex gr. *gordialis* P. et J. Thickness is 55 m.

25. Separate outcrops of clayey aleurolites and argillites. A bed of limy aleurolites is present at the top. Thickness is 10 m.

26. Rose-gray clayey fine-grained sandstones (1.5 m) with conglomerate lenses are observed at the base. Clayey rocks dominate in pebbles; sandstones and siliceous and effusive rocks are rare. Coalificated and pyritized plant remnants are present. Sandstones are rich in the bivalves *Streblopteria levis* (Lutk. et Lob.), *Streblochondria englechardti* (Eth. et Dun.), *Solemia biarmica simmetrica* (Lutk. et Lob.), and *Myonia longa* (Lutk. et Lob.). The next 23.5 meters is represented by rock debris of clayey aleurolites, argillites, and sandstones. The upper 30 m of the member comprises sandy rocks with clayey aleurolite and argillite lenses, ferruginous clay pebbles and imprints of plant trunks; interlayers of coaly-clayey rocks are observed. Thickness is 55 m.

27. Light-gray and rose-gray fine-grained sandstones with plant detritus, rare pyrite concretions, and the foraminifers *Nodosaria cuspidatila* Gerke, *Frondicularia (F.) spectata* Sossip., and *Rectoglandulina* sp. There are clayey sandstone and aleurolite beds, coaly-clayey rocks with lenses of coal, and imprints of the plants *Cordaites* ex gr. *candalepensis* (Zal.) S. Meyen and *Psygmophyllum* sp. Thickness is 30 m.

28. Dark-gray poorly sorted sandstones (15 m) with pebbles of siliceous and clayey rocks, with large limy

concretions and the rare brachiopod *Beecheria bajcurica* Tschernjak, the abundant bivalves *Streblopteria levis* (Lutk. et Lob.), *Myonia (Myonia) gibbosa* (Masl.), *Myonia* ex gr. *komiensis* (Masl.), *M. alta* (Lutk. et Lob.), *Streblochondria englechardti* (Eth. et Dun.), *S.* ex gr. *krasnoufimskiensis* (Fred.), and *Sanguinolites* sp. and indeterminable gastropods and mineralized wood. Clayey sandstones grade upwards into clayey aleurolites (25 m) with flattened limy concretions. The uppermost 25 meters is composed of clayey aleurolites, siltstones, clayey sandstones, and sandstones. The thickness of the sandstone beds increases upward to 1 m. The sandstones contain the bivalve *Sanguinolites* sp. and indeterminate plant imprints. Three thin basalt sills (0.5 m, 0.8 m, and 5 m) are observed in the upper part. The uppermost part of the upper sill shows globular jointing. Thickness is 65 m.

29. The composition is similar to that of the previous member. The lower clayey part (20 m) contains the bivalves *Modiolopsis* sp., *Strebloptria* sp., and *Polidevcia* sp. and indeterminable gastropods. An interlayering of aleurolites and sandstones is observed in next 24 m; the upper 7 meters is represented by gray fine-grained sandstones, which are replaced upward by aleurolites with lenses containing the bivalves *Streblopteria levis* (Lutk. et Lob.) and *Sanguinolites* sp. The upper part of the member is intruded by a sill (15 m) of basic rocks with cobble-stone jointing. Thickness is 51 m.

Outcrop 40 is located 300 m south of outcrop 36.

30. Dark-gray and black argillites (10) with pyrite and the bivalves *Pachymyonia bicarinata* Ast.-Urb., *Myonia (M.) komiensis* (Masl.), *Streblopteria levis* (Lutk. et Lob.), *Sanguinolites lunulatus* (Keys), and *Polidevcia* sp. are present at the base, grading upward into interlayering aleurolites and clayey aleurolites (10 m). Rose-gray inequigranular sandstones with sideritic concretions of various sizes dominate in the uppermost 20 m. The exposed thickness is 40 m.

Further description of the section is carried out at the middle part of the Zapadnyi Creek, falling into the Taimyr Lake 7 km west of Belosnezhka Creek.

Outcrop 37 is located on the left bank of the Zapadnyi Creek, 3 km southeast of outcrop 36.

The underlying beds are traced along the strike.

31. Clayey aleurolites with beds of homogeneous aleurolites that increase upward in number; sandstone beds and fine-grained sandstones occur at the top. All rocks are metamorphosed to hornfels. Thickness is 10 m.

32. Black fine-bedded coaly argillites and clayey aleurolites, intruded by two sills (0.8 m each) and a dike of basic rocks. Thickness is 10 m.

33. Dark-gray clayey aleurolites (6 m), with small gastropods and coalificated and ferruginous plant imprints. The upper 2 meters consists of gray fine-grained sandstones. Thickness is 8 m.

34. Clayey aleurolites (7 m) similar to those from member 33. Homogeneous aleurolite beds (0.1 m each) appear upward. Gray fine-grained sandstones (3 m) with pyrite concretions and indeterminable plant imprints are observed at the top. Thickness is 17 m.

35. Dark-gray clayey aleurolites and clayey sandstones (2 m), with lenses of black clay-coaly aleurolites grading upward into gray fine-grained sandstones (5 m) with coalificated and ferruginous plant imprints. There is a dolerite sill (3 m) in the middle part. Thickness of the member is 7 m.

36. Interlayering dark-gray and black argillites, clayey sandstones and, rarely, homogeneous aleurolites (5 m). Pyrite is observed throughout the section. Gray clayey sandstones with dispersed black pebbles, containing large limy lenses and the bivalves *Streblopteria levis* (Lutk. et Lob.), *S. rotunda* (Lutk. et Lob.), and *Streblochondria* cf. *sericea* (Vern.) and ferruginous plant remnants are present in the next 5-m interval. The middle 30 meters is composed of clayey aleurolites with lenses and interlayers of argillites and clayey sandstones. Gray and pinkish-gray fine-grained sandstones with large lenses of limy varieties and abundant clayey pebbles and sideritic concretions are characteristic of the uppermost 10 m. Thickness is 50 m.

The total thickness of the Tsvetochnaya Horizon is 587 m.

TATARIAN STAGE

Lower Substage

The Chernye Yary Horizon is represented here by its basal part, consisting of black argillites and clayey aleurolites, often of schistous coaly aleurolites with abundant siderite concretions of various shapes. The lower beds contain bivalves *Sanguinolites* cf. *bicarinatus* Keys. and *Streblopteria levis* (Lutk. et Lob.).

The exposed thickness is 90 m.

The age of the singled-out Permian horizons in the Taimyr region is based on brachiopods and foraminifers. The bivalves and the plant imprints play a minor role in dating.

The upper part of the *Sokolinaya Horizon* contains foraminifers, bivalves, and plant imprints. The foraminifers *Frondicularia* aff. *planilata* Gerke, *F. mica* Gerke, and *Nodosaria incelebrata* Gerke are found in the Belosnezhka Creek sequence. These fossils are known, according to G.P. Sosipatrova (Solomina, 1989), in the Nordvik region from the horizon of "diverse foraminifers" correlated to the lower part of the Kazanian Stage. However, as Sosipatrova noted, single specimens of the cited species are also found in the upper part of the Kozhevnikov lower formation, which is correlated to the Ufimian Stage, and *N. incelebrata* occurs at the base of this formation, corresponding to the Kungurian Stage. The last species extends up to the horizon of "diverse foraminifers." The bivalves *Myonia (Myonia)* ex gr. *komiensis* (Masl.), *M. (M.) gibbosa* (Masl.), and *Traberculata quadrata* (Lutk. et Lob.) are known from the Mugochan and Nuneg horizons of the Verkhoyan'e region, which are correlated to the Ufimian and Kazanian stages, respectively.

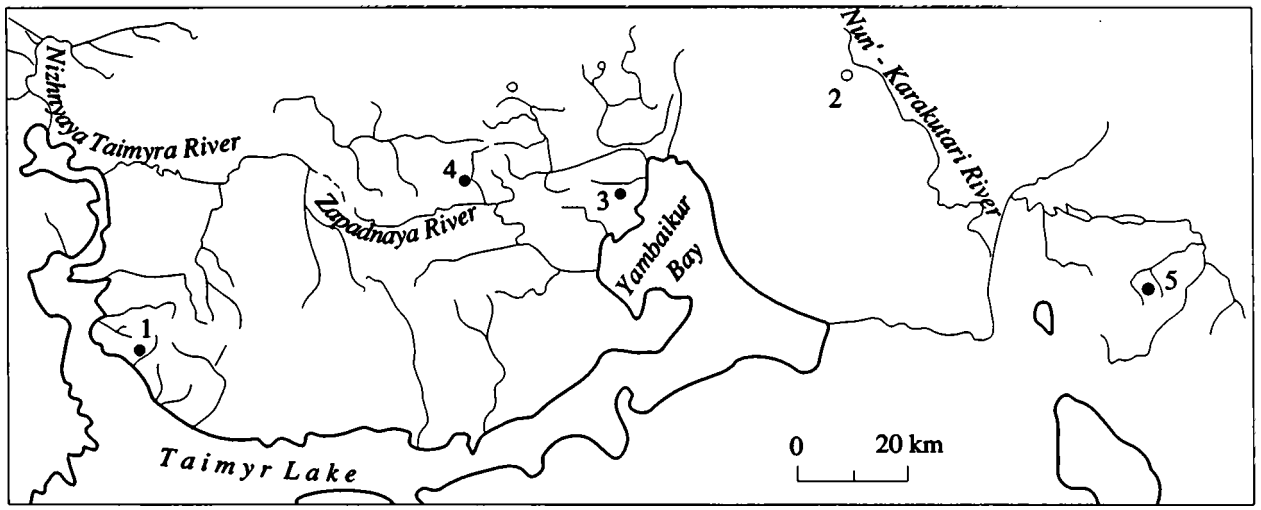

Fig. 3. *Locations of Upper Permian sections in Taimyr region. Numbers on the scheme – the outcrops of studied sections: 1 – Belosnezhka Creek, E.N. Preobrazhenskaya, 1960, R.V. Solomina, 1987; 2 – the Nun'-Karakutari River, R.V. Solomina, 1986; 3 – the Tsvetochnaya River, V.I. Ustritskii and G.P. Sosipatrova, 1963, 1983; 4 – the Snezhnyi Creek, V.I. Ustritskii, 1960; 5 – the Tulai-Kiryak Highland, Yu. A. Bogomolov, 1986.*

The Sokolinaya Horizon on the Nyun'-Karakutari River is characterized by the imprints of the plants *Koretrophyllites* sp., *Paracalamites crassus* Gorel., and *P. angustus* Such., which are typical, according to M.V. Durante, of the Balakhonka upper subformation of Kuzbass (the Lower Permian - lower part of the Upper Permian), but their presence in the lower beds of the Kol'chugino Formation (the Upper Permian) is assumed.

Noeggerathiopsis taimyrica Schwed., found on the Tsvetochnaya River, is also known, according to N.A. Shvedov, from the Balakhonka upper subformation. This allows us to correlate the beds under consideration to the upper part of the Ufimian Stage.

The Belosnezhka Horizon contains diverse brachiopods, *Orbiculoidea jangarensis* Ustr., *Strophalosia tolli* (Fred.), *Cancrinelloides yuregensis* Sol., *Terrakea* cf. *belokhini* Gan., *Rhynchopora lobjaensis* Tolm., *Olgerdia zavodowskyi* Grig., and *Bajtugania boguchanica* Sol., and foraminifers collected from its lower part on Belosnezhka Creek.

A similar assemblage of brachiopods, with the exception of *O. jangarensis* and *T. belokhini*, is found in the section on the Nyun'-Karakutari River and on the Tulai-Kiryak Highland.[1]

The brachiopods *Megousia yakutica* (Lich.), *Rhynchopora lobjaensis* Tolm., and *Licharewia* sp. (*Olgerdia* sp.)[2] are collected in the lower part of the Belosnezhka Horizon stratotype along the Tsvetochnaya River. The middle part of the horizon contains no fauna. V.I. Ustritskii (Sosipatrova, 1981) cites the fauna collected 30 km westward from Snezhnyi Creek (the Zapadnaya River) (Fig. 3). However, as these fossils do not belong to the stratotype, they are not considered here. The upper part of the horizon (the thickness is about 90 m), with the brachiopods *Orbiculoidea kolymaensis* Lich., *Cancrinelloides ochotica* (Zav.) (*C. yuregensis* Sol.), *Stepanoviella curvata* (Tolm.), *Strophalosia bajkurica* Ustr., *Rhynchopora lobjaensis* Tolm., *Licharewia schrenki* (Netsch.) (*Olgerdia zavodowskyi* Grig.), *Brachythyrina*? *sibirica* Tschernjak, and *Cleiothyridina* cf. *pectinifera* (Sow.) (*Bajtugania boguchanica* Sol.), was attributed by V.I. Ustritskii to the Baikura upper subhorizon. We believe that he defined the boundary between these two subhorizons based on lithological peculiarities: the Baikura lower subhorizon is topped by coarse-grained sandstones, and "the Baikura upper subhorizon" begins with argillites and aleurolites. As Ustritskii mentions, I.S. Gramberg found local erosion at the base of the latter. It should be added that a gap exists in the stratigraphic column 90 meters above the base. This gap masks a dislocation zone or unexposed section of undefined thickness. This might explain the thickness variations of the stratotype and the Belosnezhka Creek sections (Fig. 1). It is clear from the material cited above that most of the brachiopods from the Tsvetochnaya Horizon of V.I. Ustritskii ("the Baikura upper subhorizon") are typical of the Belosnezhka Horizon. It gives reason to draw the upper boundary of the Belosnezhka Horizon of the Tsvetochnaya River 90 m higher.

A similar brachiopod complex is widely distributed in the Verkhoyan'e region, being characteristic of the Nuneg Horizon. This complex is also known from the upper half of the Omolon Horizon of the Kolyma-Omolon region (Shvedov *et al.*, 1961) We correlate this stratigraphic level to the Lower Kazanian of the eastern

[1] The section was described by Yu.A. Bogomolov ("Aerogeologia") in 1985.

[2] From here on, redefinitions of brachiopods of R.V. Solomina are in parentheses.

European platform. Abundant remnants of the genus *Olgerdia*, which is vicarious to the Lower Kazanian genus *Licharewia*, and the presence of the genus *Bajtugania*, which is known only from the Kazanian Stage of the Urals, according to T.A. Grunt (1980), are the bases for this conclusion. This age definition is confirmed by the ammonite genera *Daubichites* and *Sverdrupites* in the Mugochan Horizon, which underlies the Nuneg Horizon. These genera are typical for the Roudian Stage, which is correlated by M.F. Bogoslovskaya to the Ufimian Stage of the Permian Chronostratigraphic Scale.

In addition to the brachiopods, the foraminifer *Nodosaria krotovi* Tscherd. is defined in the Belosnezhka Horizon on Belosnezhka Creek. According to G.P. Sosipatrova (Solomina, 1989), this species is often found in the Permian of the Polar region, together with *Pseudonodosaria ventrosa* Schleifer, which is known from the horizon of "diverse foraminifers." The bivalve *Kolymia* cf. *inoceramiformis* Lich. found in the lower part of the horizon has a wide stratigraphic range (from the Kungurian Stage up to the top of the Upper Permian). ?*Intomodesma* sp. is a typical Upper Permian genus of the Boreal realm. The leaf imprints of *Parashisoneura* aff. *sibirica* (Neub.) Radcz. are found in the upper part of the horizon. This species is known from the Il'inskoe Subformation of Kuzbass, whose age, according to the standpoint of florists, is defined as the Kazanian.

The numerous foraminifers *Nodosaria gigantea* Sossip., *N. krotovi* Tscherd., *N.* cf. *solidissima* Gerke, *Pseudonodosaria ventrosa* Schleifer, *Dentalina kalinkoi* Gerke, *Frondicularia inflata* Gerke, and *Rectoglandulina pygmaeformis* A.M.-Maclay, which are similar to those from the horizon of "diverse foraminifers," are found together with brachiopods in the 90-m-thick member on the Tsvetochnaya River. The disagreement with V.I. Ustritskii disappears, if such an understanding of the Belosnezhka Horizon range is accepted.

The Tsvetochnaya Horizon on the Belosnezhka River. The numerous foraminifers *Nodosaria incelebrata* Gerke, *N. cuspidatula* Gerke, *N. noinskii* Tscherd., *N. krotovi* Tscherd., *N. pseudoconcinna* K.M.-Maclay, *Pseudonodosaria ventrosa* Schleifer, *Lingulinella arctica* Gerke, *Rectoglandulina pigmaeformis* A.M.-Maclay, *Frondicularia mica* var. *reliqua* Gerke, *F. dilemma* Gerke, *F. planilata* Gerke, *F. pseudotriangularis* Gerke, and *Dentalina kalinkoi* Gerke, which are correlated by Sosipatrova (Sosipatrova, 1963; Solomina, 1989) to the foraminiferal assemblage from the Kozhevnikov upper formation attributed to the Kazanian Stage, are characteristic of the horizon lower part. Only the brachiopodal species *Beecheria bajkurica* Tscernjak, known from the Belosnezhka and Tsvetochnaya horizons of the Taimyr region, is found. The bivalves *Myonia (Myonia) longa* (Lutk. et Lob.) are collected in the same beds. The middle part of the horizon is enriched in the bivalves *Myonia (Myonia) longa* (Lutk. et Lob.), *M. (M.) bulkurensis* Ast.-Urb., *M. (M) komiensis* (Masl.), *M. (M.) sulkatiformis* Ast.-Urb., *Megadesmus omolonicus* Mur., *Astartella permocarbonica* (Tscherd.), and *Streblopteria levis* (Lutk. et Lob.), most of them known from the Upper Permian sediments of the Verkhoyan'e and Kolyma-Omolon regions. The bivalves *Myonia (Pachymyonia)* cf. *bicarinata* Ast.-Urb.[3] *Streblopteria rotunda* (Lutk. et Lob.), *S. levis* (Lutk. et Lob.), and *Sanguinolites lunulatus* (Keys.) are defined from the upper part of the horizon. As M.M. Astaf'eva notes, *M. (M.) longa*, *M. (M.) sulkatiformis*, *M. (P.) bicarinata*, *S. levis*, and *S. rotunda* are the most important in this assemblage. The presence of these forms allows us to correlate the host deposits to the Khal'pirki Horizon of the Verkhoyan'e region and to the Khivach Horizon of the Kolyma-Omolon region. These horizons are conditionally correlated to the Upper Kazanian Substage.

The Tsvetochnaya Horizon of the Nyun'-Karakutari River is characterized by imprints of the plants *Pecopteris* ex gr. *anthriscifolia* (Goepp.) Zal., *Cordaites* aff. *latifolius* (Neub.) S. Meyen, *C.* ex gr. *concinna* (Radcz.) S. Meyen, *C.* aff. *gracilentis* (Zal.) S. Meyen, and *C.* ex gr. *minax* (Gorel.) S. Meyen. This plant complex is represented mainly by the typical Kol'chuginian species of the Il'inskoe Subformation of Kuzbass (the Kazanian Stage approximately), according to M.V. Durante.

The regarded horizon is poorly characterized by fossils on the Tsvetochnaya River. Only the brachiopod *Beecheria bajkurica* Tschernjak is found. The bivalves are represented by *Tracia alta* Lutk. et Lob. (*Myonia alta* (Lutk. et Lob.)) and *Pseudamussium levis* Lutk. et Lob. (*Streblopteria levis* (Lutk. et Lob.)). These species are also found in the Khal'pirki Horizon of Verkhoyan'e and in the Khivach Horizon of the Kolyma-Omolon region.

The above-lying Chernye Yary Horizon of the Nyun'-Karakutari River contains, among others, abundant plant imprints of *Cordaites clercii* (Zal.), which are typical of the Kol'chugino Formation and Erunakovo Subformation of the Tatarian Stage.

The minutely described section of the Upper Permian, including the Belosnezhka and Tsvetochnaya horizons, is paleontologically better characterized than the stratotype of the Tsvetochnaya River. The two horizons are clearly distinguished: the lower brachiopodian horizon corresponds to the Belosnezhka Horizon, and the upper bivalved horizon correlates to the Tsvetochnaya Horizon. The section along Belosnezhka Creek could be recommended as a stratotype of these horizons.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 26 - 36.
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A Detailed Biostratigraphic Scheme of the Boreal Lower Triassic

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Manuscript received May 5, 1992

Abstract – A new detailed scheme of the Lower Triassic biostratigraphy is provided, consisting of 22 biostratigraphic units that were recognized using the study of sequential ammonoid assemblages from the northeastern Asia sections. The fundamental problems of global correlation within the new framework are subject to the discussion.

INTRODUCTION

The Lower Triassic biostratigraphy of Northeastern Asia is presently the most detailed, either for the Boreal Paleobiogeographic Province, or for other global regions, including that of the Tethys. Its general scheme includes 22 biostratigraphic units, or zones and subzones, which are traditionally divided into two stages: the Indian and the Olenekian (Table 2). At first glance, this result seems to be a paradox because the ammonoid taxonomic diversity in the basins of lower paleolatitudes was 2 - 3 times higher at that epoch (Kummel, 1969; Zakharov, 1975) beyond the early Indian or Greisbachian time, when the aquatories of different paleolatitudes had insufficient fauna variations.

Accordingly, one should expect a higher evolutionary rate of ammonoids in the former basins, as compared to their boreal counterparts.

A more comprehensive study of Lower Triassic ammonoids in boreal regions, as well as more favorable conditions for recognizing the succession of geological and paleontological events there, are the best explanations for the above-stated contradiction. The Lower Triassic beds in Siberia are represented mainly by claystones and siltstones, which contain interlayers and lenses of calcareous concretions bearing natural (according to E.T. Tozer, 1971) ammonoid assemblages. Quite often, the concretions contain conspecific ammonoid groups, making it possible to interprete them as fossil populations after the taphonomic-ecological appraisal. In all studied sections, the thickness of the Lower Triassic varies from tens to hundreds meters. The condensed sequences that are widespread in Tethyan regions (Tozer, 1971), which made tracing the real succession of ammonoid assemblages difficult, are absent in the boreal areas. The biostratigraphic scheme of the boreal Lower Triassic is based upon the real phylogenic branches of the endemic taxa. The phylogenetic method has a limited application for the Indian Stage, as we can trace only short specific lines within certain genera (e.g., *Otoceras, Vavilovites*), but this is the basis for subdividing the Olenekian Stage, because there are long-evolving endemic branches of various taxa: *Anaxenaspis* → *Lepiskites* → *Clypeoceratoides* (the Early Olenekian); *Xenoceltites* → *Bajarunia* → *Nordophiceras* (the Early-Late Olenekian); and *Boreoceras* → *Praesibirites* → *Parasibirites* → *Sibirites* (the Late Olenekian).

BRIEF HISTORICAL REVIEW

The first zonation of the Lower Triassic in Northeastern Asia was suggested by Yu.N. Popov (1959), who recognized five generic zones within this time interval (Table 1). The species zones were defined in the 1960s, first, for the Olenekian (Egorov and Kulikova, 1989) and then, for the Indian Stage (Vavilov, 1967). During the 1970s, the detailed stratigraphic schemes of the boreal Lower Triassic were under discussion in many publications but remained basically unchanged beyond some nomenclature modifications (Arkhipov *et al.*, 1971; Saks *et al.*, 1972; Bychkov, 1974; Dagys *et al.*, 1974; 1979). In the last decade, many more details were added to the stratigraphy of the Lower Indian (Dagys *et al.*, 1986) and the Upper Olenekian (Dagys, 1986; Dagys and Ermakova, 1988) after the revision of sequences in middle Siberia and east Verkhoyan'e. This was used as well to solve some principal problems of the last stratigraphy interval. In particular, the Lower-Middle Triassic boundary was more precisely defined, and as a result, the *tuberculatus* (= *crasseplicatus*) Zone was attributed to the Lower Anisian Stage (Dagys and Ermakova, 1984; Dagys *et al.*, 1986), while *Dieneroceras* beds (the *dieneri*, or *demokidovi* zones), which were adopted as a unit comprising the undivided Lower Olenekian upper horizons together with a lower part of the Upper Olenekian Substage, was shown to be heterogeneous (Dagys and Dagys, 1982).

The new scale under presentation (Table 2) includes all the recent achievements in the paleontology and stratigraphy of the boreal Lower Triassic. It differs from the previous frameworks by more details at the

Table 1. Evolution of ideas on the Lower Triassic zonation in northeastern Asia

Popov (1959)	Kiparisova, Popov (1964)	Vavilov (1967)	Arkhipov *et al.* (1971)	Dagys *et al.* (1979)		Dagys (1986)	
Olenekites	*tuberculatus*	*tuberculatus*	*crasseplicatus*	*spiniplicatus*		*spiniplicatus*	*subrobustus*
	spiniplicatus	*spiniplicatus*	*grambergi*				*grambergi*
Dieneroceras	*multiformis*	*multiformis*	*dieneri*	*demokidovi*	*Dieneroceras-Nordophiceras*	*demokidovi*	*contrarium*
							euomphala
					tardus	*tardus*	
Paranorites	*gracilitatis*	*gracilitatis*	*mojsisovicsi*	*hedenstroemi*		*hedenstroemi*	*kolymensis*
							hedenstroemi
Pachypropty-chites	*Pachypropty-chites*	*Proptychites*	*turgidus*	*Vavilovites*		*compressus*	
		turgidus	*strigatus*			*turgidus*	
Otoceras	*Otoceras*	*commune*	*extremum*	*nielseni*		*decipiens*	
						nielseni	
		boreale	*boreale*	*boreale*	*indigirense*	*boreale*	
					concavum	*concavum*	

infrazonal level and by a more precise fixation of stage and substage boundaries. The last point is of special interest, because the Triassic Subcommission of the International Stratigraphic Commission recommended at its meeting in Lausanne, October 1991, to use the Indian and Olenekian stages as official units of the international scale for the Lower Triassic.

BIOSTRATIGRAPHIC SCHEME

Indian Stage

In northeastern Asia, the Indian Stage is divisible into two units, as suggested in the first detailed scheme by Yu.N. Popov (1959), which included the *Otoceras* and *Pachyproptychites* Zones. His idea was supported in nearly all later publications, in which the boundary between substages was identified as the base of the *Vavilovites* beds, i.e., at the appearance level of that genus form. The only exception is the concept of Yu.V. Arkhipov *et al.* (1971), who argued in favor of three Indian substages corresponding to the *Otoceras*, *Ophiceras*, and *Vavilovites* beds, which are indeed characterized by discrete ammonoid faunas in the Siberian sequences. Imitating the subdivision principles of the oldest Triassic deposits in Canada (Tozer, 1965; 1967), the last definition of Indian substages did not become widely used.

As we noted above, the phylogenetic method is hardly applicable to the Indian Stage biostratigraphy in Siberia. There were three invasions of genetically unrelated ammonoid groups at that time. The first one was marked by the appearance of the genus *Otoceras*. The ophiceratid group, with the most characteristic genus *Tompophiceras*, is typical for the second invasion, whereas proptichids, represented solely by the *Vavilovites* genus, are the key fossils of the third. Similar distribution trends are observed in other boreal regions, but without a clear separation of *Otoceras* and *Ophiceras* faunas, because in the *boreale* Zone of Svalbard (Korchinskaya, 1986) and Greenland (Spath, 1930; 1935), ophiceratids, including the *Tompophiceras* genus, are present.

Lower Substage

The lower boundary is marked by the species of the *Otoceras* genus (*O. concavum* Tozer), appearing for the first time. At present, five biostratigraphic units can be recognized within the substage: the *Otoceras concavum*, *Otoceras boreale*, *Tompophiceras pascoei*, *Tompophiceras morpheos*, and *Wordieoceras decipiens* zones.

Otoceras concavum Zone

Hypostratotype: Bed 1 of the sequence along the Setorym Creek, Khandyga River basin, eastern Verkhoyan'e (Dagys *et al.*, 1986).

Paleontological and stratigraphic features: The zone is characterized solely by the index species.

Occurrence (within the northeastern Asia): Eastern Verkhoyan'e.

Otoceras boreale Zone

Hypostratotype: Beds 2 - 4 of the sequence along the Setorym Creek, Khandyga River basin, eastern Verkhoyan'e (Dagys *et al.*, 1986).

Paleontological and stratigraphic features: The lower boundary is marked by the appearance of the index species. Also known for certain are the *Episageceras dalailamae* Dien. and the poorly preserved xenodiscids (*Hypophiceras* and *Metophiceras*). The usually distinguished *Otoceras indigirense* Popov of this zone (Arkhipov, 1974; Dagys *et al.*, 1979) is similar morphologically to the index species. *O. domo-*

Table 2. New scheme of biostratigraphy for the northeastern Asia

Zones and Subzones			
Olenekian	upper	*Olenekites spiniplicatus*	
		Parasibirites grambergi	*Parasibirites efimovae*
			Parasibirites mixtus
			Parasibirites kolymensis
		Nordophiceras contrarium	*Praesibirites egorovi*
			Praesibirites tuberculatus
			Boreoceras lenaense
		Bajarunia euomphala	*Boreoceras apostolicum*
			Boreoceras planorbis
			Bajarunia eiekitensis
	lower	*Wasatchites tardus*	
		Lepiskites kolymensis	
		Hedenstroemia hedenstroemi	
Indian	upper	*Kingites ? korostelevi*	
		Vavilovites turgidus	*Vavilovites umbonatus*
			Vavilovites subtriangolaris
		Vavilovites sverdrupi	
	lower	*Wordieoceras decipiens*	
		Tompophiceras morpheos	
		Tompophiceras pascoei	
		Otoceras boreale	
		Otoceras concavum	

chotovi Arch. is most likely an aberrant form of the same species. Forms with a riven ventral saddle, considered as a notable feature of that species, are also known among the *O. concavum* specimens. The *Tompophiceras* genus is unknown in Siberian sequences of the zone.

Tompophiceras pascoei Zone

Stratotype: Bed 3 of the sequence along Lekeer Creek, Tompo River basin, eastern Verkhoyan'e (Dagys *et al.*, 1986).

Nomenclature: The subzone of the same appellation was originally recognized within the *Otoceras* Zone (Domokhotov, 1960), being equal to the stratigraphic interval from the first *Tompophiceras*, up to the appearance level of the *Vavilovites* genus. Later, the *Glyptophiceras extremum* (Arkhipov, 1974) and *Tompophiceras nielseni* (Dagys *et al.*, 1979) Zones were suggested to classify the *Tompophiceras* beds in northeastern Asia. Two points evidence against the last variant of zonation. First, *T. extremum* and *T. nielseni* are synonymous with the very mutable species *T. pascoei*. Second, the *pascoei* Zone under consideration has a much more limited stratigraphic range.

Paleontologic and stratigraphic features: The lower boundary is marked by the appearing genus *Tompophiceras*, whose species, *T. gracile* (Spath) and *T. pascoei* (Spath) (including *Glyptophiceras extremum* Spath, *G. subextremum* Spath, *G. nielseni* Spath, and others), are the main characteristic fossils of the zone.

Tompophiceras fastigatum Popov also originates from this zone, yet the form is a synonym of the index species. Poorly preserved smooth xenodiscids rarely occur in the zone, and ophiceratids have not been recognized so far.

Occurrence: Eastern Verkhoyan'e, Kolyma River upstream, Sea of Okhotsk coast.

Tompophiceras morpheos Zone

Stratotype: Beds 4 - 8 of the sequence along Lekeer Creek, Tompo River basin, eastern Verkhoyan'e (Dagys *et al.*, 1986).

Nomenclature: The zone is defined for the first time and corresponds to the upper part of the *nielseni*, or *extremum* Zone of preceding scales (Arkhipov, 1974; Bychkov, 1974; Dagys *et al.*, 1979).

Paleontologic and stratigraphic features: The lower boundary is marked by the appearance of index species, together with *Ophiceras* and *Wordieoceras* forms. The characteristic fauna inclùdes *Tompophiceras morpheos* (Popov), *T. bychkovi* sp. nov., *Ophiceras* ex gr. *commune* Spath, *Wordieoceras decipiens* (Spath), and *Metophiceras* ex gr. *demissum* Spath. It was argued earlier that the ammonoid assemblage of the zone includes as well some species of the *Tompophiceras* genus, such as *T. pascoei* (Spath), which are characteristic of the underlying zone (Arkhipov, 1974; Bychkov, 1974; Ermakova, 1981). A thorough revision of ammonoids provided evidence that species of the *T. pascoei* and *T. morpheos* groups are always and distinctly separated in sequences. In both the *morpheos* Zone stratotype and the Setorym Creek sequence of the Khandyga River basin, the *T. pascoei* (Spath) forms described as occurring with *Tompohpiceras morpheos* (Popov) are now classified as the new species *T. bychkovi*.

Occurrence: Eastern Verkhoyan'e, Kolyma River basin.

Wordioceras decipiens Zone

Stratotype: Bed 9 of the sequence along Lekeer Creek, Tompo River basin, eastern Vekhoyan'e (Dagys *et al.*, 1986).

Paleontologic and stratigraphic features: The lower boundary is marked by the *Tompophiceras* genus disappearance, when index species become dominant. *Wordioceras decipiens* (Spath) seems to be solely abundant within the zone; nevertheless, M.N. Vavilov (1982) reported *Ophiceras* ex gr. *leptodiscus* (Spath) as also occurring at this stratigraphic level in the Delin'ya River basin.

Occurrence: Eastern Verkhoyan'e.

Upper Substage

The lower boundary is marked by the appearing proptichids and *Vavilovites* genus dominating among the Late Indian ammonoid assemblages. The substage consists of three biostratigraphic units: the *Vavilovites*

sverdrupi Zone; the *Vavilovites turgidus* Zone, with the *Vavilovites subtriangularis* and *Vavilovites umbonatus* subzones; and the *Kingites*? *korostelevi* Zone. All zones were recognized within a single sequence of the Delin'ya River lower stream, about 30 km from the mouth (Tompo River basin, eastern Verkhoyan'e). The sequence along the Berezovyi Creek, studied by A.Yu. Egorov, S.P. Ermakova, A.G. Konstantinov, and E.S. Sobolev, is described below. Underlain by rusty-brown concretion-bearing claystones of the Lower Indian *decipiens* Zone, as indicated by abundant *Wordioceras* aff. *decipiens* (Spath), there are upsection the following beds:

1. Claystone, rusty-brown, weathered dark grey, fine flat-bedding, contains rare and small clay-calcareous concretions, occasional *Vavilovites sverdrupi* Tozer in the lower part, and poorly preserved flattened *Vavilovites* sp. in the upper. Thickness is 13 m.

2. Clayey siltstone, dark grey, distinctly flat-bedded, contains rare clay-calcareous concretions in the middle part with poorly preserved *Vavilovites* sp. Thickness is 23 m.

3. Small fault, then similar siltstone, dark grey, weathered rusty, fine flat-bedding, with rare clay-calcareous concretion interlayers, seldom *Vavilovites* sp. Thickness is 29 m.

4. Siltstone, dark grey, rusty on surface, massive, bioturbated. Thickness is 3 m.

5. Siltstone, dark grey, fine flat-bedding, contains frequent interlayers of oval calcareous concretions bearing *Vavilovites subtriangularis* Vavilov. Visible thickness is 16 m.

6. Covered interval 4 m wide, then again siltstone outcrops with numerous calcareous concretions containing *Vavilovites subtriangularis* Vavilov. Thickness is 13 m.

7. Siltstone, dark grey, mainly massive, contains dispersed calcareous concretions, *Vavilovites subtriangularis* Vavilov at the base, and *V. turgidus* (Popov) everywhere higher in the bed. Visible thickness is 15 m.

8. Small tectonic discordance, above siltstone, dark grey, distinctly flat-bedded, contains interlayers of variably sized calcareous concretions. Thickness is 20 m.

9. Siltstone, dark grey, massive, contains separate lime-concretion interlayers bearing *Vavilovites turgidis* (Popov). Thickness is 30 m.

10. Clayey siltstone, dark grey, brown on surface, includes loaflike calcareous concretions of various sizes with abundant ammonoids: *Vavilovites turgidus* (Popov), *V. compressus* (Vavilov), and *V. umbonatus* Vavilov. Thickness is 22 m.

11. Clayey siltstones, dark grey, contains small and rare flattened concretions with badly preserved ammonoids: *Kingites*? *korostelevi* Zakharov and *Metophiceras verkhoyanicum* Popov. Visible thickness is 20 m.

The higher horizons of the sequence are exposed along the right unnamed tributary of the Delin'ya River, 4 km upstream. A thick fossil-free siltstone member with fine-grained sandstone interlayers is lying here over the *Kingites*? *korostelevi* beds.

It is overlain by dark grey claystones containing *Hedenstroemia tscherskii* of the *hedenstroemi* Zone of Olenekian Stage.

In spite of the small faults cutting the described sequences, the succession of ammonoid assemblages is illustrated clearly and is suggested as a stratotype for the new scheme of the Upper Indian biostratigraphic zonation.

Vavilovites sverdrupi Zone

Stratotype: Beds 1 - 4 of the Berezovyi Creek sequence, Delin'ya River, Tompo River basin, eastern Verkhoyan'e.

Nomenclature: For northeastern Asia, this is a first-time suggested zone of given definition.

Paleontological and stratigraphic features: The lower boundary is marked by the genus *Vavilovites* appearing for the first time. The ammonoid assemblage of the zone is extremely poor. In fact, only the index species occur at this level, being clearly distinct from all younger *Vavilovites* forms by the narrow ventral side and by the thin-wall shells with flattened lateral sides.

M.N. Vavilov and Yu.D. Zakharov (1976) subdivided the *Vavilovites* genus into two subgenera. In our opinion, the range of the *Vavilovites* (*Vavilovites*) subgenus should be limited by the type species *V. sverdrupi* only. In such a case, this subgenus would occur only in the zone under discussion.

Occurrence: Eastern Verkhoyan'e, Sea of Okhotsk coast.

Vavilovites turgidus Zone

Stratotype: Beds 5 - 10 of the Berezovyi Creek sequence, Delin'ya River, Tompo River basin, eastern Verkhoyan'e.

Nomenclature: Originally, M.N. Vavilov (1967) classified the oldest Upper Indian deposits in Verkhoyan'e as the *Pachyproptychites* Zone, while the higher strata were attributed to the *Proptichites* Zone. In a recent work (Vavilov, 1982), he kept the same idea of the "thick" *Vavilovites* forms being related to lower horizons of the Upper Indian, in contrast to the "thin" forms coming out the upper strata of this time-interval. Subsequently, it was argued that there are two biostratigraphic units within the *Vavilovites* beds: the *turgidus* and *compressus* zones. In our opinion, the *Proptychites* and *compressus* zones are synonymous with the *turgidus* Zone.

Paleontologic and stratigraphic features: The lower boundary is marked by the appearance of thick subtriangular *Vavilovites* forms, which are sometimes classified as the subgenus *Vavilovites* (*Tompoproptychites*). Some authors (Vavilov, 1968; 1982; Arkhipov, 1974; Bychkov, 1974) described various *Ophiceras, Xenodiscus*, and *Proptychites* forms as occurring at this stratigraphic level, but this wasn't proved by our study. Two

units, the *subtriangularis* and *umbonatus* subzones, can be identified within the zone under question.

Occurrence: Eastern and western Verkhoyan'e, Indigirka and Kolyma riverheads.

Vavilovites subtriangularis Subzone

Stratotype: Beds 5 - 9 of the *turgidus* Zone stratotype.

Nomenclature: This is a first-time identified subzone. The index species seems to have been classified earlier as *Pachyproptychites strigatus* (Arkhipov, 1974), and our *strigatus* Zone is a synonym of the subzone.

Paleontologic and stratigraphic features: The index species guide the subzone, and the first specimens of the *V. turgidus* group appear at its level.

Occurrence: Eastern Verkhoyan'e.

Vavilovites umbonatus Subzone

Stratotype: Bed 10 of the *turgidus* Zone stratotype.

Nomenclature: This is a first-time identified subzone.

Paleontologic and stratigraphic features: The lower boundary is marked by the appearing index species. Our colleagues (Arkhipov, 1974; Vavilov, 1968; Vavilov and Zakharov, 1976) reported on a very diverse *Vavilovites* assemblage occurring in the subzone, yet most of their forms are likely synonymous with the very mutable species *Vavilovites turgidis* (Popov). Only *V. umbonatus* Vavilov and *V. compressus* Vavilov might belong to an independent species.

Occurrence: Eastern Verkhoyan'e, Indigirka and Kolyma riverheads.

Kingites? korostelevi Zone

Stratotype: Bed 11 of the Berezovyi Creek sequence, Delin'ya River, Tompo River basin, eastern Verkhoyan'e.

Nomenclature: This is a first-time identified zone. Its key species were described earlier as members of the assemblage of the underlying zone (Vavilov, 1967; Arkhipov, 1974; Bychkov, 1974).

Paleontologic and stratigraphic features: The lower boundary is marked by the final disappearance of the *Vavilovites* genus, while index species begin to occur. In addition, there are abundant xenodiscids in the zone, which are known as *Metophiceras verkhoyanicum* Popov, a species of the new genus.

Occurrence: Eastern Verkhoyan'e.

Olenekian Stage

The primary subdivision scheme of the Olenekian Stage by Yu.N. Popov (1959) included three zones of genus rank, and his idea was dominant up to the 1970s. Two substages for the Olenekian deposits in Siberia, which are quite similar to the Smithian and Spathian Stages of the standard Canadian scale by E.T. Tozer (1965; 1967), have been suggested by Yu.M. Bychkov (1974). The range of these stratigraphic units and their boundary position were more precisely defined by A.S. Dagys and A.A. Dagys (1982). According to the present interpretation, the substages correlate with two significant steps of ammonoid evolution both in boreal areas and in the basins of lower paleolatitudes (Zakharov, 1975). The wide invasion of Tethyan taxons into boreal basins took place in the Early Olenekian and resulted in the dominance of cosmopolitan genera. This correlates well with the global transgression at that time. The Late Olenekian was a phase of the extreme endemism of Ammonoidea, whose forms became very differentiated geographically (Dagys and Ermakova, 1988), supposed to have resulted from a world-wide sea level fall. The Olenekian zonation, in particular, the upper substage, is based mainly upon the phylogeny of endemic boreal taxa. As an example, the Upper Olenekian subdivision is set up on the evolutionary steps of the Sibiritidae family, reflected in the gradual transformation of most simple smooth forms into intensively ornamented taxa.

The oldest representative of that genetic line is the *Boreoceras* genus (Dagys and Ermakova, 1988), which essentially lacks the ornamentation, but demonstrates the regular change in the whorl-section contours, from round (*B. planobris*), to oval (*B. demokidovi*), and then to subrectangular with a distinctly flattened ventral side (*B. lenaense*). The *Praesibirites* genus, occurring in the higher beds, is a descendant of the above genus and preserves the whorl form of the youngest *Boreoceras*, but with appearing lateral tubercles (*P. tuberculatus*), followed by ribs (*P. egorovi*).

The *Parasibirites* genus belongs to the next step of this phylogenetic line and is peculiar because of ribbing on the ventral side and additional ventro-lateral tubercles. The species of this genus from various levels of the *grambergi* Zone demonstrate the time difference in developing ribs on the ventral side during ontogeny. The oldest *P. kolymensis* species has ventro-lateral tubercles and ventral ribbing only on the living chamber. Young specimens of this genus are practically indistinguishable from the *Praesibirites* forms. In the higher *mixtus* Subzone, species of the *Parasibirites* genus have ribs appearing on the ventral side after 5.5 - 6 whorls. Finally, the youngest species from the *efimovae* Subzone develop ribs on 3.5 - 4 whorls. The *Sibirites* genus, a terminal one for the phylogenetic line under consideration, lost its tubercles on the lateral side.

Quite often, there are transitional forms between sequential species of the described phylogeny. Such forms are also known in between species of succeeding genera (e.g., *Praesibirites* and *Parasibirites*). Having taken into account the endemism of the studied taxa, this last fact seems to be evidence that this chronostratigraphic succession reflects quite well all evolutionary steps of the Sibiritidae family. Consequently, a stratigraphic scheme based upon such data is free of sustantial hiatuses, and corresponds in details to the accuracy

of the method, because other synchronous taxa of the Late Olenekian were evolving much more slowly.

Lower Substage

The lower boundary is marked by the hedenstroemiid appearance, first, the *Hedenstroemia* genus. The substage includes the *Hedenstroemia hedenstroemi*, *Lepiskites kolymensis*, and *Wasatchites tardus* zones.

Hedenstroemia hedenstroemi Zone

Stratotype: Bed 26 (lower 105 m) of the Lekeer Creek sequence, Tompo River basin, eastern Verkhoyan'e (Dagys *et al.*, 1979).

Paleontologic and stratigraphic features: The lower boundary coincides with the substage base. The *Hedenstroemia* genus (*H. hedenstroemi* (Keys.) and *H. tcherskyi* Popov) is absolutely dominant within this zone, whereas the *Anaxenapsis* forms are less frequent. *Sakhaites* and *Kelteroceras* specimens, which become more abundant in the next zone, are reported as occurring.

Occurrence: The region from the Olenek River, on the west, to the Sea of Okhotsk coast, on the east. Without a sequence re-examination, the zone can be hardly distinguished from the *kolymensis* Zone.

Lepiskites kolymensis Zone

Stratotype: Bed 26 (upper 25 - 28 m) of the Lekeer Creek sequence, Tompo River basin, eastern Verkhoyan'e (Dagys *et al.*, 1979)

Paleontologic and stratigraphic features: The lower boundary is marked by numerous genera (*Lepiskites, Melagathiceras*, and others) coming into existence. The most diverse ammonoid assemblages occupy this zone but the only ones occurring here are representatives of the *Lepiskites, Clypeoceratoides, Meekoceras, Arctoceras, Euflemengites, Juvenites*, and *Melagathiceras* genera. Also known are last findings of the *Hedenstroemia* genus, represented by forms with complicated lobe lines (Dagys and Ermakova, 1990). *Sakhaites* and *Kelteroceras* genera occur mostly in this zone also. As indicated by ammonoid distribution, the *kolymensis* Zone seems to include two succeeding faunal assemblages. It is likely that genus *Lepiskites* takes place only in its lower part when higher horizons contain the genetically related genus *Clypeoceratoides*.

Occurrence: The *kolymensis* Zone is one of the most widespread in the Olenekian deposits of Siberia, and its counterparts are known everywhere from eastern Taimyr to northern Chukotka and the coast of the Sea of Okhotsk.

Wasatchites tardus Zone

Hypostratotype: Beds 15 and 16 of the Ken'elichi Creek sequence, Kulu River basin of the Kolyma River head (Dagys *et al.*, 1979).

Paleontologic and stratigraphic features: The appearing priontids, first of all, genera *Wasatchites* and *Anasibirites*, mark the lower zone boundary. Beyond the *Anakashmirites* genus, no other genera of *kolymensis* Zone cross this boundary. The most characteristic zonal species are representatives of three genera: *Xenoceltites* (*X. subevolutus* Spath, *X. matheri* Mathews), *Wasatchites* (*W. tardus* McLearn), and *Anasibirites* (*A. ochoticus* Bychk., *A.* ex gr. *multiformis* Welter).

Occurrence: Lena River lower course, western and eastern Verkhoyan'e, Kolyma River head, Sea of Okhotsk coast.

Upper Substage

Replacing extinct Priontidae groups, representatives of the Sibiritidae family and genus *Bajarunia* begin to occur at the substage lower boundary. Four biostratigraphic units of the substage correspond to the *Bajarunia euomphala*, *Nordophiceras contrarium*, *Parasibirites grambergi*, and *Olenekites spiniplicatus* Zones.

Bajarunia euomphala Zone

Stratotype: Beds 2 and 3 of the Toion-Uialaakh Creek sequence, Buur River basin of the Olenek River lower course (Dagys and Ermakova, 1988).

Paleontologic and stratigraphic features: The lower boundary coincides with that of the substage. The *Bajarunia* genus is the key fossil group of the zone, and the genus *Boreoceras* occurs mainly in this time interval. Here as well, the genera *Koninckitoides* and *Pseudosageceras* make their appearance. The zone is subdivided into three subzones in accordance with distribution of the *Bajarunia* and *Boreoceras* forms.

Occurrence: The zone has been reported as taking place in most structural provinces of northeastern Asia (Dagys *et al.*, 1979), but this opinion needs revising because of difficulties in recognition of the *Boreoceras* and *Bajarunia* species.

Bajarunia eiekitensis Subzone

Stratotype: Bed 2 of the zone stratotype (Dagys and Ermakova, 1988).

Paleontologic and stratigraphic features: The lower boundary corresponds to that of the zone. Rare ammonoids are represented only by index species.

Occurrence: Northern Midsiberia (Olenek River lower course).

Boreoceras planorbis Subzone

Stratotype: Bed 3 (lower part) of the zone stratotype (Dagys and Ermakova, 1988).

Nomenclature: The zone was first defined by A.S. Dagys and S.P. Ermakova (1984), but in a wider range than it is now. Additional study revealed two sequential and discrete units with their index species

occurrence limited by the lower portion of the original range.

Paleontologic and stratigraphic features: The appearing *Boreoceras* genus indicates the position of the subzone lower boundary. In addition to index species, evolute *Bajarunia* forms (*B. alexandri* Dagys) are quite characteristic of this unit, but other species (*B. taimyrensis* Dagys, *Pseudosageceras* sp.) are rare.

Occurrence: Eastern Taimyr for certain, Lena and Olenek rivers interfluve.

Boreoceras apostolicum Subzone

Stratotype: Bed 3 (upper part) of the zone stratotype (Dagys and Ermakova, 1988).

Nomenclature: This is a first-time distinguished subzone.

Paleontologic and stratigraphic features: The appearance of an index species and *Boreoceras demokidovi* Kipar., which resembles it, is a mark of the subzone lower boundary. Both occur only at this stratigraphic level, and no *Bajarunia euomphala* (Keys.) species have been noted for certain outside. The first forms of *Koninckitoides posterius* (Popov) appear in the subzone.

Occurrence: Characteristic species of the zone were recognized in a wide area between Taimyr and the coast of the Sea of Okhotsk, but this data needs to be verified.

Nordophiceras contrarium Zone

Stratotype: A sequence of the Tulai-Kiryaka Upland, eastern Taimyr. It was described by Dagys *et al.* (1989), but it now needs slight correction in the classification of certain species, as is done in the following description. The sequence starts with the East-Taimyr Formation, containing *Bajarunia euomphala* (Keys.) and *Boreoceras* aff. *apostolicum* (Smith). It is overlain by the Ystannakh Formation, whose section is most representative of the *contrarium* Zone, being divisible into four units:

1. Alternating clay- and siltstones, dark grey, contains subordinate beds of tuffaceous silt- and sandstones, together with frequent horizons of calcareous concretions bearing *Boreoceras lenaense* (Dagys et Konst.). Thickness is 82 m.

2. Clayey siltstone, dark grey, with repeated concretion interlayers and ammonoids: *Nordophiceras contrarium* (Popov), *Praesibirites tuberculatus* (Dagys et Konst.), and *P.* aff. *egorovi* Dagys et Erm. (transitional forms). Thickness is 32 m.

3. Siltstone, dark grey, fine flat-bedding, contains flattened concretions with *Nordophiceras* aff. *contrarium* (Popov), *Praesibirites egorovi* Dagys et Erm. Thickness is 37 m.

4. Irregularly interlayering claystones and clayey siltstones. Thickness is 20 m.

These rocks are overlain by the fossil-free tuffaceous sandstones of the Pribrezhnyi Formation.

Paleontologic and stratigraphic features: The lower boundary is distinguishable by the appearing genus *Nordophiceras*. The zone contains the last specimens of *Boreoceras* (*B. lenaense*), and genus *Praesibirites* occurs solely within this interval. *Koninckitoides* and *Pseudosageceras* forms are quite usual at the basal level. The whole unit consists of three subzones.

Occurrence: Eastern Taimyr, Lena and Olenek rivers interfluve, Kharaulakh, eastern Verkhoyan'e.

Boreoceras lenaense Subzone

Stratotype: Bed 1 of the presented sequence.

Paleontologic and stratigraphic features: The lower boundary is that of the zone. The subzone bounds the occurrence of *Boreoceras* forms with flattened ventral sides (*B. lenaense, B. mirabile*) and the genus *Evenites*. Last specimens of *Koninckitoides posterius* (Popov) have been recognized in it.

Occurrence: Eastern Taimyr, Lena and Olenek rivers interfluve, Kharaulakh, eastern Verkhoyan'e.

Praesibirites tuberculatus Subzone

Stratotype: Bed 2 of the presented sequence.

Nomenclature: The subzone was originally defined by A.S. Dagys and S.P. Ermakova (1988), but without delimitation from the *egorovi* Subzone. There are transitional forms between the index species of the *tuberculatus* and *egorovi* subzones, but typical *P. egorovi* occurs exclusively in the upper part of the *contrarium* Zone.

Paleontologic and stratigraphic features: The appearing *Praesibirites* genus serves as an indicator of the subzone lower boundary. Beyond the index species, there is abundant *Nordophiceras contrarium* (Popov).

Occurrence: Eastern Taimyr, Kharaulakh.

Praesibirites egorovi Subzone

Stratotype: Bed 3 of the presented sequence.

Nomenclature: This is a first-time established zone.

Paleontologic and stratigraphic features: The lower boundary corresponds to the appearance level of index species. In addition, there are also known forms of *Nordophiceras* aff. *contrarium* (Popov), which very likely belong to a new species.

Occurrence: Eastern Taimyr.

Praesibirites grambergi Zone

Stratotype: The most representative zone sequence was described by A.Yu. Egorov and L.I. Kulikova (1989) along the Chernokhrebetnaya River in eastern Taimyr. There are several ammonoid-bearing levels traceable within Member 4 of the Ystannakh Formation, consisting of clayey siltstones with interlayers of calcareous concretions:

a) Lower 10 m – *Bajarunia euomphala* (Keys.) and others (the *euomphala* Zone);

b) Upsection 10 - 30 m – *Nordophiceras contrarium* (Popov), *Boreoceras lenaense* (Dagys et Konst.);

c) Upsection 30 - 50 m – *Praesibirites tuberculatus* (Dagys et Konst.) (the *contrarium* Zone);

d) Upsection 50 - 55 m – *Parasibirites rariaculeatus* (Popov), *P. kolymensis* Bytschk. (the *grambergi* Zone, *kolymensis* Subzone);

e) Upper 55 - 70 m – *Parasibirites grambergi* (Popov), *P. mixtus* (Popov), *P. subpretiosus* (Popov), and *Sibirites elegans* Dagys et Erm. in the lower part (the *grambergi* Subzone), then *Parasibirites efimovae* Bytschk. and *Sibirites subpretiosus* (Mojs.) in the upper part (the *efimovae* Subzone).

Paleontologic and stratigraphic features: The lower boundary is marked by the appearing genus *Parasibirites*. The *Nordophiceras, Boreomeekoceras, Sibirites, Olenekoceras*, and *Subolenekites* genera are also abundant in the zone, which can be divided into three new subzones.

Occurrence: Eastern Taimyr, Lena and Olenek rivers interfluve, Kharaulakh, eastern Yakutia, Kolyma River head.

Parasibirites kolymensis Subzone

Stratotype: (see the zone stratotype).

Paleontologic and stratigraphic features: The lower boundary coincides with that of the zone. Besides index species, *Nordophiceras menense* (Popov) and *Subolenekites pilaticus* Tozer are characteristic forms of the zone.

Occurrence: Eastern Taimyr, Lena and Olenek rivers interfluve, Kolyma River head.

Parasibirites mixtus Subzone

Paleontologic and stratigraphic features: Such species as *Parasibirites grambergi* (Popov), *P. mixtus* (Popov), and *P. subpretiosus* (Popov) mark the lower boundary of the zone. The first *Sibirites* forms (*S. elegans* Dagys et Erm.) appear at this level.

Occurrence: Eastern Taimyr, Lena and Olenek rivers interfluve, Kharaulakh, eastern Yakutia.

Parasibirites efimovae Subzone

Paleontologic and stratigraphic features: The appearance level of index species and forms of the *Olenekoceras* and *Boreomeekoceras* genera is the lower zone boundary. The zone assemblage includes: *Sibirites pretiosus* (Mojs.), *Boreomeekoceras keyserlingi* (Mojs.), *Olenekoceras levigatum* Dagys et Erm., and *Nordophiceras kazakovi* Dagys et Erm.

Occurrence: Eastern Taimyr, Lena and Olenek rivers interfluve, Kolyma River head.

Olenekites spiniplicatus Zone

Stratotype: The Mengilyakh Creek sequence, Olenek River lower course (Egorov and Kulikova, 1989).

Paleontologic and stratigraphic features: In indicating this zone, *Olenekites, Keyserlingites, Pseudosvalbardiceras, Timoceras*, and *Prosphingites* genera mark its lower boundary by their appearance. Many species of other genera, such as *Nordophiceras karpinskii* (Mojs.), *N. schmidti* (Mojs.), *Sibirites eichwaldi* (Keys.), and others, are also typical for the zone. It seems to include a number of sequential assemblages. In particular, it is worth mentioning that genus *Prosphingites* is related only to the upper horizons of the zone.

Occurrence: Eastern Taimyr, Lena and Olenek rivers interfluve, Kharaulakh, eastern Verkhoyan'e, Kolyma River head.

CORRELATION

1. Boreal regions. The correlation problems of the Lower Triassic zonation in Siberia and other boreal regions have been discussed in several recent works (Dagys and Tozer, 1989; Weitschat and Dagys, 1989); here, only certain more precise correlatives (Table 3) will be under consideration.

The Boreal Zones in the eastern and western parts of the Boreal Province are not equivalents. Those of Arctic Canada, Greenland, and Svalbard contain various *Tompophiceras* forms from the *T. pascoei* group and can be correlated with the *boreale* and *Pascoei* zones in Siberia. Therefore, only the *morpheos* Zone of Siberia, bearing true *Ophiceras* forms, is analogous to the *commune* Zone in Canada.

The Upper Indian Substage is an extremely complicated stratigraphic interval to be correlated. In Siberia, there is no indication of strata similar to the *candidus* Zone in Canada, and so far, this gap in paleontological record of the Upper Indian basal horizons corresponds to a single hiatus into the stratigraphic scheme of the Lower Triassic in northeastern Asia. Contrary to the previous idea, one may suggest that the *Vavilovites* beds of Siberia and Canada are not entirely synchronous. In the western part of the Boreal Province, there are only horizons bearing the first relatively thin *Vavilovites* forms with a narrow ventral side (the *sverdrupi* Zone), yet thick species of the *Vavilovites* (*Tompoproptychites*) subgenus, which are the most diverse and widely distributed in Siberia, are absolutely unknown. Consequently, correlatives of the *turgidus* Zone seem to be absent in the western boreal regions. No equivalents of the *korostelevi* Zone, terminal in the Indian Stage, have been recognized there, either.

The correlation of the Lower Olenekian zones is undoubtedly established because of the ammonoid cosmopolitism at this time interval. However, the *hedenstroemi* Zone occurs widely only in Siberia, and in the western regions, it has been recognized only in a single locality of Arctic Canada (Dagys and Tozer, 1989).

Zonation schemes of the Upper Olenekian in eastern and western sections of the Boreal Province are incomparable concerning their details. Only two biostratigraphic units, with a significant hiatus at the substage base, have been defined in the best studied sequences in Canada, as opposed to the ten zones defined in Siberia. The stratigraphic levels of the *euomphala* and *contrarium* zones are totally unknown in all western regions, though the sequence lithology does not evidence any gaps in sedimentation. The next hiatus comprises a significant part of the *grambergi* Zone, because the type species of the *pilaticus* Zone in Canada occurs solely in the *kolymensis* Subzone in Siberia. Only beds correlative with the *spiniplicatus* Zone have been noted within the entire Upper Olenekian Substage of Svalbard (Weitschat and Dagys, 1989).

2. Tethyan regions. The Lower Triassic zonation of the Boreal and Tethyan regions can be directly correlated only for the periods of maximum transgressions. In fact, we currently know only two such stratigraphic data. One corresponds to the basal level of the Indian Stage and comprises the *Otoceras* beds, which locally occur in the Tethyan area, just along the southern domains adjacent to the Gondwanaland. Under these circumstances, detailed correlations are contraversial. E.T. Tozer (1981) considers the Otoceras beds of the Boreal and Tethyan localities as synchronous in general, though we believe that the *boreale* Zone can be correlated for certain only with the *woodvardi* Zone, because the *concavum* Zone equivalents have not been established in the lower paleolatitudes (Dagys and Dagys, 1986) (Table 3).

The Lower Olenekian is the second stratigraphic datum. Its *tardus* Zone can be unmistakably recognized all over the world, and this correlation level is proved by the conodont determinations (Dagys and Dagys, 1986), because the *Neogondolella milleri* Zone is the precise equivalent of the *tardus* Zone, again having the worldwide distribution. The *kolymensis* Zone correlatives are also quite distinct because of the widely occurring cosmopolitan genera, yet outside the southern Tethys area, where units analogous to the *tardus* Zone are underlain by strata correlating only with the *hedenstroemi* Zone, they are not so evident (Dagys and Ermakova, 1990). The last zone correlatives are less clear and can be sufficiently recognized mainly within the Gondwanaland areas of the Tethys (Dagys and Ermakova, 1990).

So far, there is no data even for a rough correlation between the Upper Indian zones of the Boreal and Tethyan regions. The detailed relations between the Upper Olenekian biostratigraphic schemes remain problematic as well. B. Kummel (1966; 1968; 1969),

Таблица 3. Correlation between zonation schemes of the Lower Triassic

Stage	Substage	Northeastern Asia		Canada		Svalbard	Western USA	Primor'e	Himalayas	Salt Range
Olenekian	upper	*spiniplicatus*		*subrobustus*		*subrobustus*	*mckelvei*	*multiformis*		*pakistanum*
		grambergi	*efimovae*	Spathian			*parisianus*	*insignis*		*Tirolites-Columbites*
			mixtus							
			kolymensis		*pilaticus*					
		contrarium	*egorovi*							
			tuberculatum							
			lenaense							
		euomphala	*apostolicum*							
			planorbis				*harti*	*Tirolites*		
			eiekitensis							
	lower	*tardus*		Smithian	*tardus*	*tardus*	*Anasibirites*	*nevolini*	*Ps. himaica*	*pluriformis*
		kolymensis			*romunduri*	*romunduri*	*gracilitatis*	*bosphorense*		*gracilitatis?*
		hedenstroemi			*hedenstroemi*				*rohilla*	*flemingianus*
Indian	upper	*korostelevi*		Dienerian				*subdharmus*	Layers with *Meekoceras*	*rotundatus-volutus*
		turgidus								
		sverdrupi			*sverdrupi*	*sverdrupi*				*frequens*
					candidus					
	lower	*decipiens*		Griesbachian	*strigatus*	*rosenkrantzi*			*tibeticum*	*connectens*
		morphaos			*commune*					
		pascoi			*boreale*				*woodwardi*	
		boreale								
		concavum			*concavum*					

L.D. Kiparisova (1961), and Yu.D. Zakharov (1968) have described many boreal genera of the Upper Olenekian in sequences of lower paleolatitudes, yet these determinations seem to be mostly incorrect (Dagys and Ermakova, 1984).

The real *Bajarunia* and *Olenekoceras* were collected in Primor'e from *Tirolites* and *Columbites* beds, respectively (Zakharov, 1968; Zakharov and Rybalka, 1987). Species of *Olenekoceras* (*Keyserlingites stephensoni*), *Bajarunia* (*Prionolobus jacksoni*), and *Boreoceras* (*Dieneroceras apostolicum*) were recognized in the *Columbites* beds of western USA.

According to this data, the *Tirolites* beds can be correlated with a lower part of the *euomphala* Zone in Siberia. The *Columbites* beds (the *parisianus* and *insignis* zones) are comparable with the upper subzone of the *euomphala* Zone and with the range of the *contrarium* and *grambergi* zones, because the first *Olenekoceras* forms appear in Siberia just at the top of *grambergi* Zone. Such correlations make it possible to assume that the *parisianus* Zone of western USA and the *insignis* Zone of Primor'e, rich in ammonoid genera, are based upon the "unnatural assemblages" of E.T. Tozer (1971) and most likely include several sequential faunas. Being terminal in the Olenekian Stage of northeastern Asia, the *spiniplicatus* Zone can be well correlated with the *subrobustus* Zone of Canada, which contains a number of genera (*Zenoites, Isculitoides*, and *Procarnites*) characteristic to the sub-*Columbites* beds of the Tethyan regions. It was assumed, however, that this zone of the boreal Triassic can be placed at the level of the *haugi* Zone (Silberling and Tozer, 1968; Tozer, 1971; 1981; Bucher, 1989), which was originally recognized within the Anisian Stage in western USA. Characteristic Middle Triassic assemblages (Acrochordiceratidae and Longobarditidae) have been recognized in the last zone in association with peculiar forms, such as "*Tirolites*" *pacificus*, which belongs, according to H. Bucher (1989), to the genus *Keyserlingites*, being similar to *K. subrobustus* (Mojs.), the most typical species of the *spiniplicatus* Zone. Y. Wang (1984) believes that the genus *Keyserlingites* occurrs in both the Upper Olenekian and the Lower Anisian and has been occupying boreal basins during the Early Triassic and Tethyan regions during the next stage. Such transmigration of the genus is doubtful, and the problem is rather caused by the fact that Anisian forms, including those from the *haugi* Zone of western USA, belong to the genus *Durgaites*, which is a heterochronous homeomorph of the genus *Keyserlingites* (Dagys, 1988).

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 38 - 45.
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Evolution of the biosphere

On the Age of Mammalian Fauna from the Karabulak Formation of the Kalmakpai River (Zaisan Depression, Eastern Kazakhstan)

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Manuscript received July 2, 1992

Abstract – The study of fossil mammalian fauna from the Kalmakpai River locality (Karabulak Formation deposits) indicates that this fauna must be referred to the Turolian of the European stratigraphic scale, i.e, to the terminal Miocene, and not to the Ruscinian, as was previously supposed. The bone bed is correlated with the beginning of polarity Chron 5 of the magnetochronologic scale.

INTRODUCTION

The study of the Cenozoic deposits in the Zaisan Depression has a long history. It was most intensive at the end of the 1950s and the beginning of the 1960s. B.A. Borisov (1963) generalized the geological data on this region according to the results of the geological survey performed by the Leningrad group of the Altaian Expedition (East Kazakhstan Geological Department) from 1956 to 1961. The suggested stratigraphic scheme included many subdivisions, characterized paleontologically using different groups of the biota.

The Neogene section along the Kalmakpai River was subdivided into the middle Miocene Zaisan Formation, the middle-upper Miocene Sarybulak, the lower-middle Pliocene Kalmakpai, and the middle-upper Pliocene Karabulak Formations.[1] In 1960, hipparion remains were found in the middle part of the Karabulak Formation, as well as enamel fragments of a hypsodont tooth, originally considered as belonging to an elephant. V.E. Garutt and L.K. Gabuniya determined the age of the remnants as middle-late Pliocene. Afterwards, the remnants proved to be tooth fragments of *Sinotherium* (*Elasmotheriini*). Nevertheless, the first identification led to the overstating of the age up to the middle-late Pliocene (Borisov, 1963).

[1] These ages are corrected according to the Interregional Stratigraphic Committee scheme (1956), in which the Pontian corresponds to the lower Pliocene.

In 1961, 1966, and 1968, scholars of the Paleontological Institute, Russian Academy of Sciences, made excavations that provided essential osteological data.

In 1966, V.I. Zhegallo published a preliminary list of mammals. He determined the middle Pliocene age of the Kalmakpai fauna (Karabulak Formation), considering that the underlying Kalmakpai sediments corresponded to the lower Pliocene Pavlodar Formation, whereas the fauna was somewhat more progressive than that of Pavlodar. This age was also adopted later (Vangengeim *et al.*, 1972); however, it was stated, "According to the generic spectrum, the large mammalian fauna in this stage of evolution was rather similar to the preceding Pavlodar one" (p. 59).

The excavated remnants were thoroughly studied by different investigators. E.L. Dmitrieva (1977) described from this locality the *Gazella dorcadoides* Schlosser and referred it to the lower-middle Pliocene. V.I. Zhegallo identified two hipparion species *Hipparion hippidiodus* Sefve and *H. elegans* W. Grom. and referred the fauna from this locality to the Ruscinian, the beginning of the middle Pliocene (Kimmerian), considering its age as post-Pavlodar. A.Ya. Godina (1979) described a new subgenus and species of the giraffe *Palaeotragus* (*Yuorlovia*) *asiaticus* and dated it as the Early Pliocene. Yu.A. Semenov (1989) described the ictithere remnants as *Hyaenictitherium hyaenoides orlovi*. M.V. Sotnikova (1992) identified a new species *Machairodus kurteni* and *Adcrocuta eximia* (Roth et Wagner).

Table 1. Comparison of the Kalmakpai fauna with Turolian faunas of Pavlodar and China

Kalmakpai	Pavlodar	Turolian faunas of China
Vormela sp.	–	–
Adcrocuta eximia (Poth et Wagner)	+	+
Hyaenictitherium hyaenoides orlovi Semenov	*H. venator*	*H. hyaenoides*
Martes palaeosinensis Zdansky	+	+
Melinae gen. indet.	–	–
Plesiogulo crassa Teilhard	cf.	+
Indarctos sp.	–	+
Machairodus kurteni Sotnikova sp. n.	*M. irtyschensis*	*M. gr. tingii-palanderi*
Hipparion hippidiodus Sefve	–	+
Hipparion elegans W. Gromova	+	–
Chilotherium sp.	+	+
Sinotherium zaisanense Bajshashov	sp.	*S. lagrelii*
Cervavitus novorossiae Chomenko	–	+
Procapreolus latifrons Schlosser	sp.	+
Gazella dorcadoides Schlosser	+	+
Tragocerus sp.	+	+
Palaeotragus (Yuorlovia) asiaticus Godina	+	+
Samotherium cf. irtyshense Godina	*S. irtyshense*	3 species

Additionally, the collection of the Paleontological Institute yields remnants of *Chilotherium* sp., *Sinotherium* sp., *Cervinae*, *Tragocerus* sp., and *Vormela* sp.

In 1981 - 1983, the excavations in this locality were conducted by paleontologists from the Institute of Zoology, Kazakh Academy of Sciences (Tleuberdina, 1988). The investigations supplemented the previous list of mammals with *Melinae* gen.?, *Indarctos* sp., *Plesiogulo crassa* Teilh., *Cervavitus* sp., and *Procapreolus* sp. B.U. Baishashov (1986) referred the *Sinotherium* remains as a new species *Sinotherium zaisanensis* Baish. It appears that *Hipparion houfenense* was erroneously mentioned in this list with the reference to V.I. Zhegallo's (1978) paper. These authors determined the age of the Kalmakpai fauna as the Kimmerian and assigned it to the Ruscinian.

In the regional overview on Eastern Kazakhstan (Stratigrafiya SSSR, 1986), the Kalmakpai Formation was referred to the lower Pliocene (Pontian), whereas the Karabulak Formation, including the bone bed, to the middle Pliocene (Kimmerian). In the same volume, L.K. Gabuniya (p. 325) erroneously assigned the mammal remnants from the Karabulak Formation to the Kalmakpai Formation and dated them as the end of the Early or beginning of the middle Pliocene (Pontian-Kimmerian).

Therefore, the paleontologists' opinions on the age of the Kalmakpai mammalian fauna vary. They place it in the Pontian, i.e., the lower Pliocene, at the lower-middle Pliocene boundary, or finally, in the Kimmerian (the middle Pliocene), but the majority of investigators, however, correlate it with the West European Ruscinian.

It should be remembered that according to the European stratigraphic scale of continental deposits, the second half of the Miocene is divided into the Vallesian and Turolian stages, corresponding to the upper Serravalian, Tortonian, and Messinian, whereas the Pliocene stages are the Ruscinian and Villafrancian (lower and middle). The upper Villafrancian is referred to the Quaternary.

During the last decade, intensive study of the Neogene mammalian fauna throughout almost all the Palaearctic revealed general regularities of its evolution and allowed the stratigraphic range of many taxa to be defined more precisely. In accordance with modern views on the Neogene history of the Palaearctic fauna, we had to revise our opinion on the place of Kalmakpai fauna in the general stratigraphic system.

COMPOSITION AND STRATIGRAPHIC DISTRIBUTION OF TAXA FROM THE KALMAKPAI FAUNA

The list of taxa, identified from the Kalmakpai locality, is represented in Table 1, and their stratigraphic distribution is shown in Fig. 1.

Martes palaeosinensis Zdansky was recorded in the Paote Formation, Shansi Province in China. Yu.A. Orlov (1941) described *Martes* sp., from the Pavlodar locality, with features characteristic of *M. palaeosinensis*. A similar species *M. andersoni* is known from the Ertemte locality that is dated as the terminal Turolian and the beginning of the Ruscinian. E. Anderson (1970) considered these forms to be synonymous. Therefore,

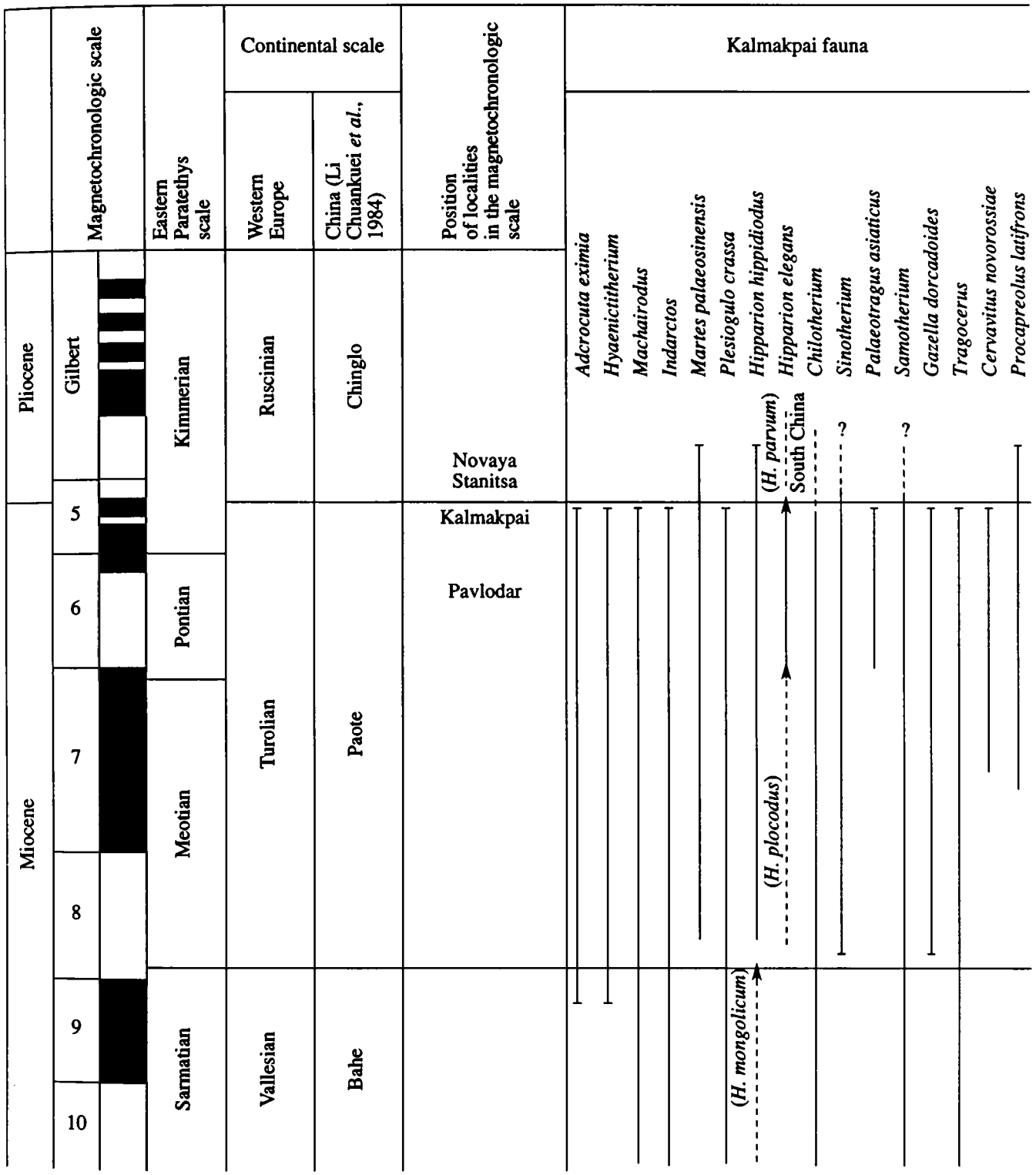

Fig. 1. *Stratigraphic distribution of mammals from the Kalmakpai River locality and its position in the magnetochronologic scale.*

the stratigraphic range of *M. palaeosinensis* covers the Turolian and the beginning of the Ruscinian.

Fossil remnants of the genus *Vormela* occur very rarely. Its lower stratigraphic limit is unknown.

The Neogene glutton *Plesiogulo crassa* Teilhard was widely distributed in the Chinese Turolian (the Paote Formation). It was considered to be a characteristic form of Central Asia. Recently, however, the occurrence of the Neogene glutton was recorded in Europe. *P. crassa* was recorded in the upper Turolian Cherevichnoe locality in the northern Black Sea region (Korotkevich and Semenov, 1975), the lower Turolian Vathylakkos locality in Greece (Koufos, 1982), and the upper Messinian Baccinello locality in Italy (Rook, 1990). Remains of this form were also found in the sediments of the Dhok-Pathan Formation in India. Yu.A. Orlov (1941) recorded *Plesiogulo brachygnathus* in the Pavlodar. B. Kurten (1970), revised the genus *Plesiogulo* and ascribed this species to *P. crassa*. Thorough study shows the Pavlodar form to be rather small and closer to *P. brachygnathus*, according to its dimensions. Nevertheless, its main morphological features are similar to those of *P. crassa*, and thus we assign it to *P. cf. crassa*.

Adcrocuta eximia (Roth et Wagner) is known in Eurasia from the end of the Vallesian (second half of the MN 10 zone) to the terminal Turolian. It was especially widely distributed in the second half of the Turolian from China to Western Europe and Libya. It was also recorded in Pakistan.

Hyaenictitherium hyaenoides (Zdansky) was an important faunal component in the Chinese Turolian and most likely inhabited Turkey (Semenov, 1989). From the Kalmakpai locality, Yu.A. Semenov (1989) described a new subspecies *H. hyaenoides orlovi*, which "differs from the nominal subspecies by less massive canines, relatively narrower premolars, and weaker reduction of last molars" (p. 126). It is not clear whether this is a geographical difference or it indicates younger age.

In Eastern Europe, this genus is represented by another species *H. venator* Semenov that is known since the end of the Late Sarmatian (Late Vallesian) and was especially characteristic of the second half of the Meotian (Middle Turolian). This genus was not recorded after the Turolian.

The genus *Machairodus* is known in Eurasia since the beginning of the Vallesian. A new species *M. kurteni* Sotnikova was recorded in the Kalmakpai locality. The arrangement of the incisors of this form has some special features peculiar to the later genus of serrate-toothed cats *Homotherium*. This makes the Kalmakpai form different from the typical Turolian sabre-toothed cats of the *M. giganteus* group, and it can be considered as the latest representative of the genus *Machairodus*.

Two hipparion species were recorded, *Hipparion hippidiodus* Sefve and *H. elegans* W. Gromova. *H. hippidiodus* was absolutely dominant. This species appeared as a result of the Turolian adaptive radiation of *Hipparion mongolicum* Schlosser, which was characteristic of the Vallesian analogues in Central Asia. *Hipparion hippidiodus* is attributed to the unique group of hipparions without a fore-eye socket. Its first representatives migrated to Europe and Africa from Central Asia in the Turolian and initiated the appearance of *H. platigenis* and *H. turcanense*. Late forms of *H. hippidiodus* are typical for the Chinese Paote Formation (the Turolian), but they were also recorded in its Asian analogues of the Early Ruscinian, e.g., the Leijiahe locality in China (Li Chuankuei, 1984).

Hipparion elegans is very rare, in contrast to the Pavlodar locality, where it dominates. It is a characteristic form of the Late Turolian, being phylogenetically connected with the Early Turolian *H. plocodus*, and in the beginning of the Pliocene, it initiated a new evolutionary stage of *H. parvum*, the smallest among Asian hipparions.

The rhinoceroses are represented in Kalmakpai by two genera, the *Sinotherium* and the *Chilotheruim*. Both groups have an Asian origin. They have roots in the Oligocene, at the beginning of the habitat redistribution among *Rhinocerotidae*, which replaced hyracodonts.

Sinotherium is known only in the Asian part of the Palaearctic. It existed for certain in the Turolian, Late Miocene. A long hiatus between *Sinotherium* and its probable descendant elasmothere, known since the end of the Pliocene, does not offer the possibility of valid stratigraphic interpretation of the *Sinotherium* find in the Kalmakpai locality. Nevertheless, it is important that *Sinotherium*, as a constant component of fossil assemblage, occurs with one exception only in the Miocene and is not known for certain stratigraphically higher than the Ertemte level (Inner Mongolia).

The form established in Kalmakpai was considered by B.U. Baishashov (1986) as a new species *Sinotherium zaisanensis*, but it has no reliable difference from *S. lagrelii*, a common species in the Turolian faunas of China.

The genus *Chilotherium* appeared in South Asia in the Agenian. In the Palaearctic, it was common in Vallesian and Turolian faunas. It is a rather morphologically stable genus. Its earliest and latest species sufficiently differ in the degree of hypsodonty, but the morphological distinctions between the synchronous or near synchronous species are fairly relative. The latest chilothere remnants are known from the base of the Pliocene in the localities of South China (Gansu) and Western Mongolia (Chirgis Nuur 2).

Deer are represented by *Cervavitus novorossiae* Chomenko and *Procapreolus latifrons* Schlosser. The former was widely distributed in the first half of the Turolian among the northern Black Sea region faunas. The species was recorded in the Grebeniki, Tarakliya, and other localities. In China, it is known from the Paote fauna of the Henan and Shansi Provinces, e.g., the localities Paote, Yushe, zone I, and others (Teilhard, 1937; Zdansky, 1925).

The distribution of *P. latifrons* appears to have been limited to Central Asia. It was recorded in the Olan-Chorea and Ertemte localities of Inner Mongolia and China (Schlosser, 1924), which yielded fauna transitional between the Turolian and Ruscinian. In China, this form was also common in the earlier faunas of the Paote Formation. It was also recorded in the Turolian of Mongolia (Dzachso-Chairchan 1).

Gazella dorcadoides Schlosser is a characteristic form of the Turolian fauna in China. It is known in the localities of Chirgis Nuur 2 in Mongolia (member A, Turolian) and Ortok in Kirgizia. A form close to *G. dorcadoides* was recorded in Pavlodar (Dmitrieva, 1977).

The Kalmakpai fauna yields two giraffe genera, *Palaeotragus* and *Samotherium*. The genus *Palaeotragus* occurred extremely widely in the Eurasian hipparion faunas, whereas the distribution area of the subgenus *Yuorlovia* was mainly limited by the Asian part of the Palaearctic. The most eastern occurrence of the giraffe, which can possibly be related to this subgenus, was recorded in the eastern Azov region (Godina and Baigusheva, 1985). Except for the Kalmakpai locality, *Palaeotragus* (*Yuorlovia*) *asiatica* Godina is known from the Pavlodar, Ortok, and Chinese Turolian faunas (Godina, 1979). *Samotherium* was also common enough in the Turolian of Eurasia. Reports on its remains in younger localities require a more accurate approach.

Representatives of Tragocerini appearing in the Middle Miocene, namely, in the Astaracian (Korotkevich,1981), flourished in the Late and middle Turolian. Toward the end of the Turolian, their number decreased greatly, and in the Ruscinian localities they were not found.

TAPHONOMY OF THE LOCALITY AND ECOLOGICAL CHARACTERISTICS OF THE FAUNA

The Kalmakpai fossil assemblage is enclosed in a pluvial lens inside the proluvial-deluvial sedimentary complex. The proluvial channel is not clearly seen in the section, and its thalweg can be recorded only by erosional pockets in the middle of the sequence. The flood running along the thalweg was constant enough to produce the channel-line facies, represented by pure gravel, and the flood-plain facies, composed of poorly sorted deposits containing remnants of small vertebrates (Ochotonidae indet.). Therefore, the fossil assemblage was formed near a more or less steady flood that provided the riverside vegetation resources sufficient for the numerous chilotheres, as follows from the excavation data. At the same time, the fossil assemblage position at the piedmont of the proluvial-deluvial train and the complete absence of alluvial facies in the Karabulak Formation clearly indicate the low humidity of extravalley biotopes of that time.

The dominance of *Hipparion hippidiodus*, an inhabitant of the open landscapes and an obligatory herbivorous form, as well as the sparse occurrence of the browsing *H. elegans* and the presence of gazelle with higher hypsodonty, also confirm the conclusions about the xerophytic landscape during the time of the Kalmakpai fauna.

PALEOZOOGEOGRAPHIC CHARACTERISTICS OF THE FAUNA

Thorough study of the Kalmakpai fauna taxonomic spectrum indicates its striking resemblance to the Turolian hipparion fauna of Northern China (Table 1). In 1903, M. Schlosser distinguished in Northern China two paleozoogeographical provinces, the forest area in the south and the steppe in the north. Each of them, together with common forms, yields certain peculiar mammalian species. B. Kurten (1952) called them the "gaudry" (with a typical representative being *Gazella gaudryi*) and "dorcadoides" (with *Gazella dorcadoides*) faunas, respectively. The former inhabited the territories of Henan, Hebei, and the southeastern Shansi provinces, and the latter, the Gansu and northwestern Shansi. The transitional zone (part of Shansi and Shensi), with mixed, according to Kurten, fauna, occurred between those paleozoogeographical provinces.

The genera and some species of the Kalmakpai locality, equivalent to the North Chinese fauna forms, were distributed in the Kurten's fauna-provinces as follows: *Sinotherium* occurred only in the dorcadoides fauna; *Samotherium*, *Chilotherium*, *Tragoceros*, *Indarctos*, and *Plesiogulo*, in the dorcadoides and transitional faunas; *Hipparion hippidiodus* and *Gazella dorcadoides* had the same distribution; *Hyaenictitherium hyaenoides* was present in the dorcadoides and transitional faunas and occurred in the single Hung-Kou locality of the gaudry fauna.

Adcrocuta eximia (= *Hyaena variabilis* in Chinese lists) was most characteristic in the dorcadoides, rarely occurring in the transitional and being found only once in the gaudry fauna. The genus *Procapreolus* is known from the transitional and gaudry fauna localities.

Palaeotragus was abundant in the Palaearctic hipparion faunas, whereas the distribution area of subgenus *Yuorlovia* occurred only in the eastern Palaearctic. *P. (Y.) asiaticus* was found in the dorcadoides and transitional faunas (Godina, 1979).

GEOLOGICAL AGE OF THE KALMAKPAI FAUNA

As cited above, the Kalmakpai fauna is evidently recorded in the same paleozoogeographical province as the Northern Chinese hipparion dorcadoides fauna. The majority of the Kalmakpai forms does not exceed the Turolian bounds (Fig. 1).

In the former USSR territory, the Kalmakpai fauna is most similar to the Gusinyi Perelet fossil assemblage near the Pavlodar (Pavlodar Formation deposits, Table 1). These faunas have the following common

forms: *Adcrocuta eximia, Hipparion elegans, Chilotherium, Sinotherium, Samotherium, Palaeotragus* (*Yuorlovia*) *asiaticus, Gazella dorcadoides, Tragocerus,* and *Procapreolus.*

The genus *Hyaenictitherium* is represented in the Gusinyi Perelet locality by another species, *H. venator.* The Turolian age of the Pavlodar fauna is beyond any doubt. Nevertheless, the complete synchronization of the Kalmakpai and Gusinyi Perelet faunas appears to be invalid. For example, the Kalmakpaian gazelle is more hypsodont than that of the Pavlodar fauna. This indicates a more arid environment and, most likely, a younger age.

Giraffes also have some differences. The Kalmakpaian *Palaeotragus* is more long-legged. However, these differences most likely can lie within the intraspecific variability or be connected with somewhat different areas of inhabitation. Teeth of the Kalmakpaian *Samotherium* have narrower crowns than those of the Pavlodar form.

The dominance of *Hipparion hippidiodus* in the latter locality, characteristic of the terminal Turolian faunas, may serve as indirect evidence of the age difference between the Gusinyi Perelet and Kalmakpai localities. Accordingly, we can correlate fauna from the Karabulak Formation of the Kalmakpai River with the very end of the Turolian and consider it to be younger than the Gusinyi Perelet fauna.

The paleomagnetic record also indicates the age difference of these localities. In the Gusinyi Perelet section, the bone beds are associated with reversely magnetized deposits that correlate with polarity Chron 6 and the Pontian of the Eastern Paratethys.

According to V.L. Yakhimovich[2] (oral communication, Yakhimovich *et al.*, 1990), deposits at the top of the Karabulak Formation show a reversed polarity; the bone bed is associated with the normal polarity zone (about 7 - 10 m thick); the lower, greatest part of the upper subdivision of the Karabulak Formation (about 17 - 20 m) is reversely magnetized; the lower subdivision is associated with the zone of predominantly normal polarity. Deposits at the base of this formation are reversely magnetized.

Two zones of reversed and normal polarity were distinguished in the Sarybulak Formation sediments.

Yakhimovich correlates the Karabulak Formation with the Gilbert paleomagnetic epoch, attributing the bone bed to the Cochiti event and the upper part of Chron 5; the Kalmakpai Formation, with the beginning of Chron 5 and Chrons 6, 7, and 8; and the Sarybulak Formation, with Chrons 12 and 13. She supposes the presence of a considerable hiatus between the Kalmakpai and Sarybulak Formations.

According to the data of M.A. Pevzner and E.A. Vangengeim (1986), the Turolian-Ruscinian boundary in the European continental stratigraphic scale must be placed within Chron 5 and almost coincides with the Miocene-Pliocene boundary. In the Eastern Paratethys stratigraphic scale, it is placed in the lower Kimmerian.

Basing on the pre-Ruscinian age of the Kalmakpai fauna, the paleomagnetic section on the Kalmakpai River can be interpreted in the following way (downward from the top).

Normally magnetized deposits with the bone bed, most likely correspond to the lower part of Chron 5 (basal Kimmerian); the reversed polarity zone, underlying the bone bed, to Chron 6, i.e., to the Pontian and Gusinyi Perelet beds of the Pavlodar Formation; normally magnetized deposits of the lower subdivision of the Karabulak Formation and top of the Kalmakpai Formation, to Chron 7, i.e., the Late Meotian. The Kalmakpai Formation sediments correlate with the beginning of Chron 7, Chrons 8 and 9, and the end of Chron 10 and correspond to the lower half of the Meotian and Late Sarmatian. The Sarybulak Formation can be correlated with Chrons 10 and 11, i.e., with the middle and part of the Early Sarmatian. The characteristic Astaracian forms *Anchitherium* sp. and *Amphicyon* sp., from the base of the Sarybulak Formation, agree well with this dating, suggested by L.K. Gabuniya ("Stratigrafiya ...," 1986). Regarding such interpretation of the paleomagnetic record, we consider that the section on the Kalmakpai River is free of large hiatuses, in particular, that between the Kalmakpai and Sarybulak Formations, which was supposed by V.L. Yakhimovich to correspond to the Early-middle Sarmatian. Accordingly, the stratigraphic volume of the Kalmakpai Formation becomes wider, whereas that of the Sarybulak Formation decreases.

CONCLUSIONS

The study of the taxonomic composition of the fossil mammalian fauna from the Kalmakpai River locality and correlation with the paleomagnetic record enables us to draw the following conclusions.

The Kalmakpai fauna is referred to the Turolian, i.e., we should place it in the terminal Miocene of the European stratigraphic scale. In the magnetochronologic scale, it corresponds to the lower part of Chron 5 and is stratigraphically younger than the fauna of the Gusinyi Perelet locality, which corresponds partly to the level of the Pavlodar Formation. The latter is correlated with Chron 6. The Novaya Stanitsa locality of the Irtysh River, with the lower Ruscinian fauna (Fig. 1), takes place at the boundary between polarity Chron 5 and the Gilbert epoch (Vangengeim *et al.*, 1984; Zykin *et al.*, 1989) and has a stratigraphically higher place than the Kalmakpai locality.

In Central Europe, the upper Turolian Baltavar locality in Hungary has the same position in the magnetochronologic scale as the Kalmakpai fauna. Deposits bearing fossil mammalian remnants (beds with *Unio wetzleri*) show a normal polarity and overlie the Pon-

[2] The authors are very grateful to V.L. Yakhimovich for permission to use her unpublished data on the paleomagnetic record of the Kalmakpaian section.

tian sediments that contain mollusks (Pevzner and Vangengeim, 1986). In Western Europe, the Messinian localities Baccinello and Brizighella in Italy may also be analogous, according to the carnivore mammal species (Rook, 1990).

The most essential change in the previous dating of the Kalmakpai fauna concerns its correlation with the European continental scale: it should be referred not to the Ruscinian, but to the Turolian. As for correlation with the Eastern Paratethys stratigraphic scale, the position of bone beds can now be defined more accurately: they correspond to the basal Kimmerian only.

The Kalmakpai fauna is not only one of the latest Turolian faunas in Asia, the richest among the precisely dated fossil assemblages of that period, but also marks the final Turolian stage of the last precrisis hipparion faunas preceding the Ruscinian.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 45 - 54.
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Genetic Classification of Late Quaternary Deposits of the Black Sea Eastern Margin

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Manuscript received May 15, 1991

Abstract – The lithological composition of the Late Quaternary deposits in the continental slope and adjacent part of the Black Sea deep-water depression was studied. Genetic types of sediments of the canyons, their submarine cone, and intercanyon slope were distinguished near the Caucasian coast. A relationship of the section's structure with the sea floor microrelief elements was established. The dynamics of sedimentation was traced back for certain periods of the Quaternary.

The eastern margin of the Black Sea is a geologically and geomorphologically complex part of the basin. Here, the folded structures of the Greater Caucuses submerge below the sea floor at an oblique angle. The complex relief of the mountainous land, with numerous dissections of river valleys, is reflected distinctively in the relief of the sea floor. The latter is marked by a narrow shelf and steep continental slope rugged with canyons. At a depth of more than 1500 m, the slope turns into a continental rise where submarine fans are formed. An abrupt shift of sedimentation conditions, both along and across the continental slope, results in complex facial patterns of sediments and a variety of their genetic types.

Most of the data available now on sediments under discussion concerns the regional aspects of modern and quaternary sedimentation. A few studies touch the problem of sedimentation in the canyons of the Caucasian coast of the Black sea (Zenkovich, 1976; Leont'ev and Saf'yanov, 1972; Mel'nikov *et al.*, 1979; Saf'yanov, 1975; Saf'yanov and Pykhov, 1977; Strakhov, 1954; Shimkus *et al.*, 1975). They consider the structure and lithodynamics of the canyon systems of the Inguri, Kodori, Rioni, and other rivers.

At present, there is an urgent necessity to create a general model of sedimentation on the continental slope. This model should explain how sedimentary patterns of various series depend on the transportation and accumulation of the sediments, i.e., on their genesis. In turn, the origin is conditioned by both regional and local conditions. The former are the width of the shelf, the contrast of the land and subaqueous relief, the epeirogenic movements, the geochemical composition of the water, the biological productivity, and so on. The latter are the forms of the slope microrelief, the degree of remoteness of the sedimentation zone from the shore, the steepness of canyon slopes, the configuration of their bottoms, and the height and form of the banks and watershed crests.

Here we attempted to show the influence of as many sedimentation factors as possible on the formation of sedimentary series. We also tried to reveal the dependence of textural, structural, mineralogical, and geochemical properties of a sediment on its genesis. At the same time, we compared the sedimentation patterns in two closely situated canyon systems, differing in the intensity of depositional processes and in the composition of terrigenous clastic material. The analysis undertaken allowed us to shed light on some aspects of the paleogeography and geologic evolution of this part of the continental slope in the Quaternary.

The Black Sea is a favorable object to study sedimentation processes in canyons and their submarine fans. First, this is stipulated by the distinct expression and relatively small dimensions of these structures. Second, unlike for many other seas and the ocean, there is a simple and reliable scale of the Late Quaternary deposits subdivision, based on the specific compositional features and the chronological succession of sediments. The scale enables us to reconstruct a history of the basin and easily correlate the core sections.

We studied sediments sampled within two canyon systems. The first is situated between Gelendzhik and Tuapse and the second is in the Shakhe River region (Figs. 1, 2). In the first case, canyons begin at the shelf edge and join at a depth of about 1500 - 1800 m, forming a single submarine fan. A canyon of the Shakhe River begins near its mouth, crosses the slope at almost a right angle, and widens at a depth of some 1700 - 1900 m. Sediment cores were sampled in various parts of the canyons and fans across their strike. Echo sounding was used to establish the morphological features of the canyons.

In the sedimentary section, the Novochernomorian (New Black Sea), Drevnechernomorian (Old Black Sea), Novoeuxinian (New Euxinic), Karangat, and Chauda deposits have been established (Strakhov, 1954; Shimkus *et al.*, 1975).

The Novochernomorian deposits are distinguished by a large quantity of interlayers and lenses formed of coccolithes *Emiliania huxleyi* (Lohm). They are made up of the interbedding of several-centimeter-thick terrigenous argillaceous muds and several-millimeter- to a few-centimeter-thick coccolithic-sapropelic pares. The Drevnechernomorian deposits are distinct in their particular banded pattern, formed by layers of sapropelic mud. The Novoeuxinian deposits are represented by calcareous clayey muds with interlayers and lenses, enriched with hydrotroilite, and with brakish-water microfauna. The Karangat deposits bear the coccolithes *Gephyrocaspa caribbeania* Borud. et Hay and are made up of clayey and calcareous muds up to several centimeters in thickness. The Chauda beds are established on the basis of the index molluscs of the *Didacna* and *Dreissena* genera (determinations of E.V. Babak). We also refer compact fossil-free clays to them.

Our data enabled us to subdivide the studied interval of the section into separate beds, members, and associations. We also attempted to reveal the formation mode of these lithological units, i.e., the genetic types of sediments.

To establish this or that genetic type, that is, to determine the mode of transportation and deposition of sedimentary material, we used the structural features: layer boundary patterns, differences in granulometric composition, and interrelations of lithological units. The composition of the sediment is conditioned by the provenance, geochemical conditions, and biological productivity of the basin. The composition can vary considerably within a single genetic type, because it is independent of the transportation and accumulation mode of the sedimentary material. The bed thickness of the deposits having the same genesis can vary in a similar manner. However, the relative thickness remains quite constant for the whole sedimentation period in a given region.

Erosion particularly is to be noted among the processes that have formed the sequences of canyon systems. Erosion is registered by an absence of some stratigraphic intervals in sections, exposures of bedrocks in some parts of canyons, and redeposition. The presence of autochthonous sediments, i.e., those formed by the washout and redeposition of muds on the same continental slope, is also the result of erosional processes.

The defined genetic types of sediments, corresponding to particular deposition patterns, can be divided into three groups.

1. Sediments formed by gravitational processes. These are deposits of landslides and various autokinetic flows: mudflows (pebble-gravel and gravel-sandy), clastic, grain, and turbidity flows. Sediments of the group characterize the degree of the canyon system activity and the steepness of the slope. They accumulate in thalwegs and on the banks of canyons. An exception is presented by analogues of distal turbidites. They are formed on the watershed (intercanyon) elevations as a result of a "splashing out" of the main stream material. Their composition is similar to distal turbidites of

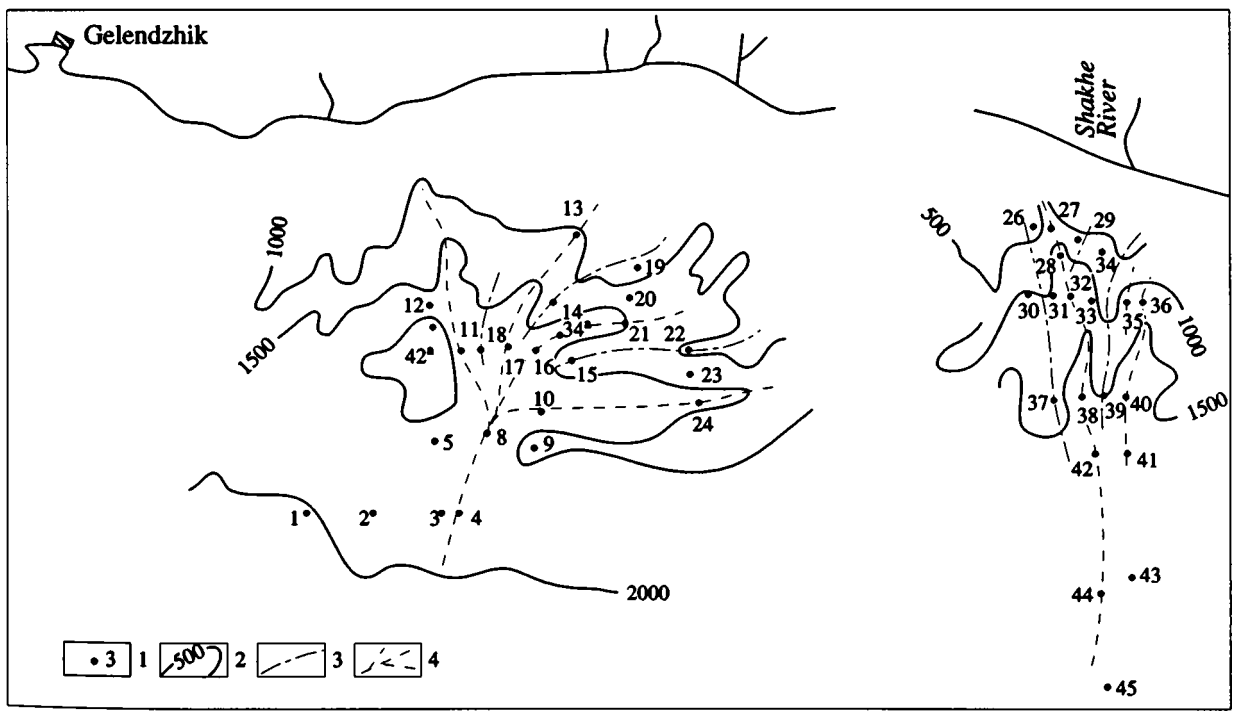
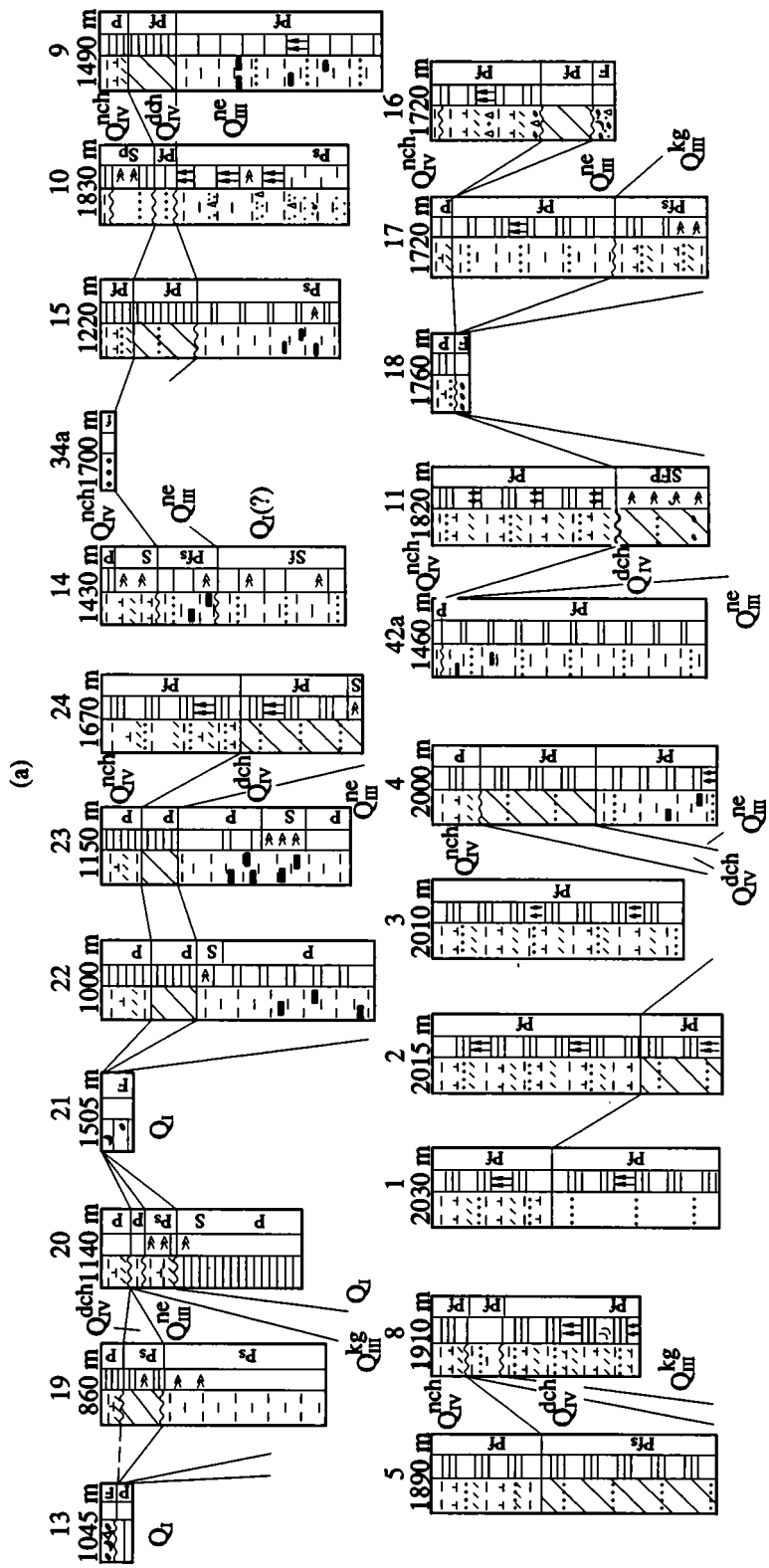


Fig. 1. *A sketch of the studied area. 1 – stations of sediment sampling; 2 – isobaths, m; 3 – axes of watersheds; 4 – axes of canyons and channels.*


(a)
1 2030 m
2 2015 m
3 2010 m
4 2000 m
5 1890 m
8 1910 m
9 1490 m
10 1830 m
11 1820 m
13 1045 m
14 1430 m
15 1220 m
16 1720 m
17 1720 m
18 1760 m
19 860 m
20 1140 m
21 1505 m
22 1000 m
23 1150 m
24 1670 m
34a 1700 m
42a 1460 m
Q(?)

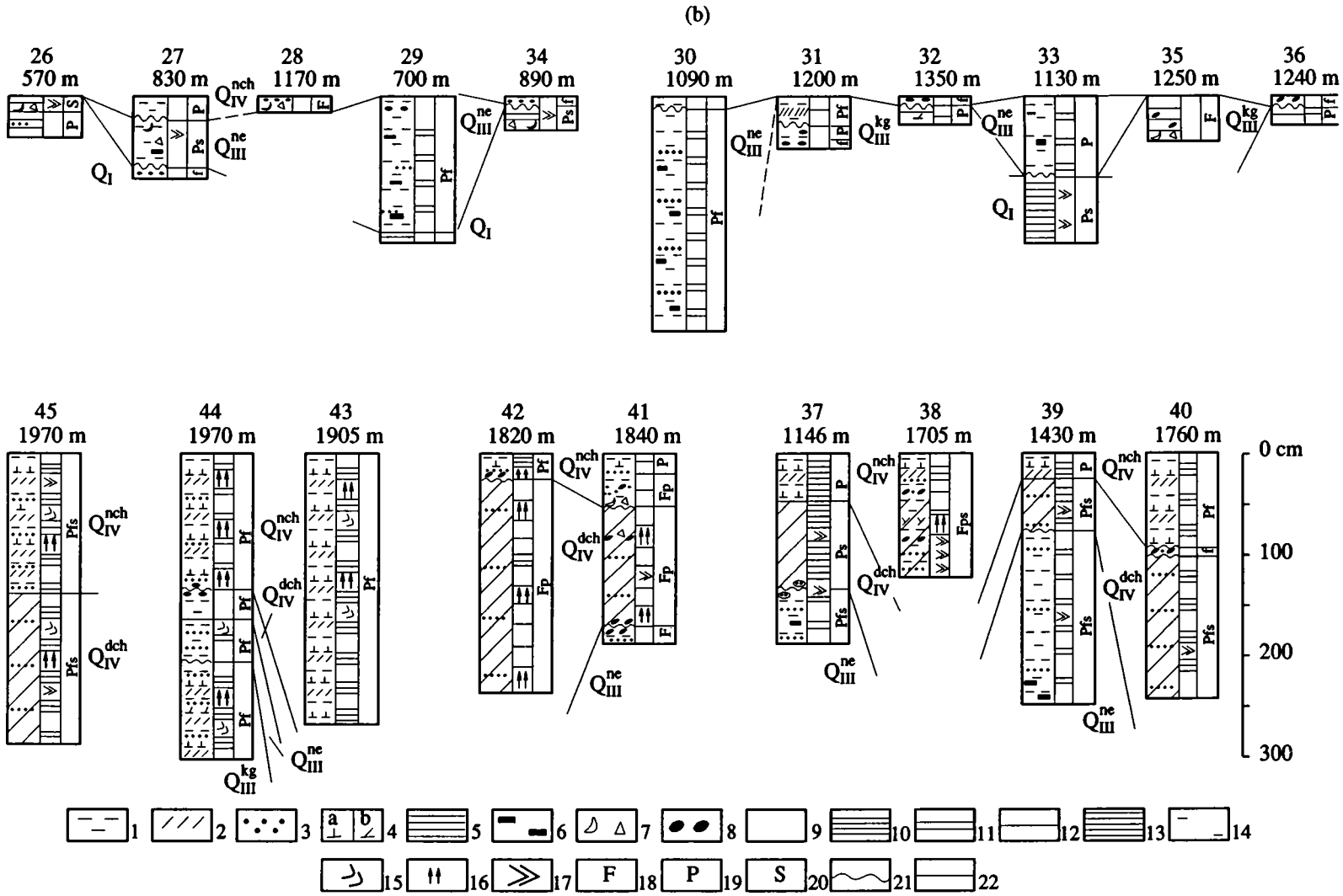


Fig. 2. *Sections of sediments: (a) Gelendzhik-Tuapse region; (b) Shakhe River region; Q_{IV}^{nch}. Novochernomorian beds; Q_{IV}^{dch}. Drevnechernomorian beds; Q_{III}^{ne}. Novoeuxinian beds; Q_{III}^{kg}. Karangat beds; Q_I. Chauda beds; 1 – clayey muds; 2 – sapropel; 3 – sand; 4 – calcareous mud (a – biogenic; b – chemogenic); 5 – clay; 6 – hydrotroilite; 7 – shells and detritus; 8 – pebble and gravel; Textures: 9 – massive; 10 – parallely banded with elements of microlamination; 11 – parallel-banded; 12 – parallel-stratified; 13 – microlaminated; 14 – lens-like; 15 – cross-bedded; 16 – graded bedding; 17 – slumping; Genetic groups of sediments: 18 – pelagic; 19 – of autokinetic flows; 20 – landslide; 21 – erosional boundaries; 22 – flat boundaries.*

submarine cones. The composition of the sediments of this group depends on local or coastal sources of the material supply. The alternation of various types of sediments in the section indicates a change in a canyon regime. Spatially, it reflects a distance of the sampling point from the shore.

2. Sediments formed by the gradual sedimentation of particles from the pelagic suspension (particle by particle). This group comprises pelagic sediments constituting an integral genetic type, but with considerably variable composition. These sediments reflect either periods of weakening of the gravity-induced processes in canyons, or they indicate areas of undisturbed sedimentation under sluggish hydrodynamic conditions. The composition of these deposits varies in different stratigraphic levels. Lateral compositional changes are more rare. The latter reflect the evolution of the geochemical, thermal, and biological conditions in the basin.

3. The third group includes deposits of bottom currents. They reflect the hydrodynamic activity of the deep water masses.

Below, we give the criteria to define these genetic types, followed by a brief description of the sediments.

Slide deposits can be defined by the typical structural features of slumping folds, by the lens-like imbedding of older deposits into younger ones, and by the plunging contacts with the enclosing muds. We have found indications of basin slumpings only. They are confined to canyon banks and to thalwegs. In certain sections, signs of slumpings as a rule predate erosional processes, reflecting their intensification in the course of time.

Deposits of *debris flows* are characterized by poor sorting of the clastic material and a high proportion of clayey matrix (>50%). They often have erosional lower and abrupt upper boundaries. Clastic material consists of redeposited products from the coastal zone, shell detritus, and pebbles of poorly lithified and reworked underlying muds. The rolling degree of the lithoclasts is fairly good, beach-like. Mud pebbles are flattened and lens-like. Gravel and sand usually compose unsorted sediments (mixtites) with massive structure. Pebbles of intrabasin muds are mostly oriented parallel to the bedding plane within layers. In a single core (station 10, 100 - 120 cm), we came across a more complex distribution of mud pebbles. It was chaotic in the lower part of the bed, oriented across bedding in the middle part, and flat-bedded above. Deposits of mudflow debris, enriched in shelly detritus, contain mainly arenaceous clastic material and a smaller amount of clayey matrix (about 40%). A slight gradation can be seen here as a result of an increasing concentration of shell fragments and a decreasing concentration of floating pebbles toward the upper parts of the layer. Sometimes we observed an enrichment of the middle part of the layer in gravel material and an inverse gradation with shell detritus at the base and gravel material at the top of the layer. Deposits of debris flows are confined to thalwegs of the middle and lower parts of the canyons. They range from 5 to 40 cm in thickness, depending on the size of the clastic material. Deposits of what we term *clast flows* should be referred to the same group. They differ from the deposits described above in a lower content of clayey matrix.

Deposits of *grain flows* are distinct in a low content of clayey matrix (<20%) and in the absence of shelly detritus. Their sandy clastic material is fairly well sorted. Bed boundaries are distinct, with the lower one often being erosional, sometimes with clay lenses at the base. The thickness of the grain flow deposits varies within the limits of the first dozens of cm. These sediments are confined to thalwegs of the middle and lower parts of the canyons and associate closely with mud- and clast flow deposits.

Deposits of *turbidity flows* are the most widespread among gravitationally formed sediments. They are distinguishable by the presence of graded bedding with all or some elements of the Bouma sequence (Bouma, 1962). There are distinct, often erosional lower boundaries and gradual transitions into the overlying pelagic sediments. The terrigenous fraction is finely psammitic or silty with a low content of clayey matrix (20 - 30%). The occurrence of turbidites with all the Bouma sequence elements is quite rare. More often, these sediments have the elements "a–b" (proximal turbidites) or "b–c–d" and "c–d–e" (distal turbidites). Convolute bedding in the "c" element has been registered only once. Normally this element contains an interlayer with the finest, ranging within fractions of a millimeter, lense-like bedding of silty material. Distal turbidites contain varieties formed by the reworking of earlier deposited muds. Sapropelic particles, tiny coccolith clusters, and diatoms play the role of clastic material here. Depositions of these sediments apparently occurred at the expense of erosion of the underlying muds somewhere up the slope, when they were disturbed and then deposited from the suspended flow. The turbidite nature of this sediment is indicated by graded bedding on a macro- and microscopic scale. The amount of sapropelic material decreases upwards, and the base of the layer is enriched with dispersed small lenses of coccolithic material. At the top of layer there is a gradual transition into pure mud, sometimes bearing scattered plant remnants. Thin sections show a constant presence of also graded silty material in this type of turbidite. We encountered turbidites in all microforms of the slope relief. Their proximal varieties are associated with thalwegs of the canyons' lower parts, with submarine cones, the canyons' banks, and watershed crests. Turbidites are absent at some isolated elevations only.

Pelagic sediments are accumulated as a result of a gradual sedimentation of particles from the pelagic suspension. The composition of the deposits and the ratio of terrigenous to biogenic components and the organic matter in them vary in different periods of the basin evolution. The composition of the pelagic muds presents a foundation for the stratigraphic subdivision of

the Late Quaternary sediments of the Black Sea (see above). They are characterized by massive structures in homogeneous clays with traces of diagenetic alteration and by extremely fine flat-bedding. The latter, being formed by the seasonal deposition of a material, reflects an alternation of clayey, carbonate coccolith, and organic matter laminas. Pelagic muds make up members ranging in thickness from several centimeters to several meters, with the thickest registered at the foot of the continental slope in submarine fans. In particular sections, they often form regular paragenetic associations with sediments of the turbidity fan muddy tails or with deposits of weak discrete bottom currents. In the first case, there are transitions across flat-bedded sediments from turbidites into pelagic deposits ("d–e" elements of the Bouma sequence). In the second case, there is an alternation of pelagic oozes with thin interbeds of silt and silty clay deposits, sometimes having a cross-sinuous structure.

A specific type of sediment is associated with the activity of *bottom currents*. We attribute to this type thin silty layers having a flat boundary with the underlying deposits. Their upper boundary is often uneven, bearing swells and sometimes, distinct ripple marks. It can also bear wedging out interlayers with inclined upper surfaces. A cross-bedding in these layers can be detected in thin sections being commonly emphasized by plant remnants or mica flakes. More rarely, the cross-bedded structures can be seen visually. Some sediment types occurring in the Novoeuxinian oozes apparently belong to this group of bottom current deposits. These are characterized by inclined parallel lamina of hydrotroilite (stations 19, 29) changing to a horizontal position up the section. Under the microscope, the hydrotroilite lamina are enriched in fine silty clastic material. The thickness of these members can amount to 40 cm. Cross-bedding in thin sections is sometimes expressed in layer-by-layer aggregations of pyrite clasts, grains, crystals, and globules.

DISTRIBUTION OF DIFFERENT SEDIMENT TYPES

The echo sounding survey has revealed a number of relief microforms within the continental slope and rise. They are classified as canyon thalwegs, canyon banks with terraces and escarpments, intercanyon crests, thalweg bars and other positive relief forms, and channels and interchannel areas of submarine fans.

Canyon bottoms and thalwegs. Our data shows that thalwegs morphology depends on the distance from the shore and on the angle of the canyon's erosional incision into the slope basement (profile of incision). The other factors are the bottom form and width of the canyon.

Thalwegs of the upper part of the slope were mainly the area of transportational and erosional processes in the Novochernomorian and, possibly, during the Drevnechernomorian time. In some canyons, even older Novoeuxinian and Karangat deposits are affected by erosion. Sometimes all of the Novochernomorian sediments are eroded. These features most intensively manifested in the eastern canyon system (Fig. 2b). Here, down to the 1500 m isobath, the Novochernomorian deposits are absent on channel banks and elevations, whereas in thalwegs, only the Karangat and Chauda deposits are exposed.

The core sampled in the middle part of the slope in the eastern canyon system (station 38) revealed the Novochernomorian deposits in the thalweg zone. They consist of pebble and gravel material deposited by debris flows and include slide members (10 to 20 cm thick). The upper part of the core bears several 2- to 5-cm-thick interlayers of coccolith-sapropelic pelagic sediments.

In the western area, thalweg deposits were sampled from the same depths at stations 11, 12, 16, and 17 (Fig. 2a). Sediments of station 17 characterize the most intensive activity period within the canyon. At the base of this section, a slide member of the Karangat deposits overlain by undisturbed pelagic sediments was obtained. The Novoeuxinian muds are represented by an alternation of the pelagic sediments (5 - 25 cm thick) and proximal turbidites ("a–b", "a–d" elements of Bouma sequence, 5 - 45 cm thick). The element "c" is usually represented in turbidites by thin, lens-like silty interlayers that sometimes undulate and bifurcate. The bottoms of turbidite beds are enriched in sand and shell detritus. Higher, the Novochernomorian deposits lie over the erosional surface. They consist of coccolith-sapropelic muds with lenses of more compact Novoeuxinian muds composed of fine-grained shell detritus and silt. The Drevnechernomorian deposits are absent. Muds of this age are also absent from station 11. Here, the Novoeuxinian sediments are represented by a 60-cm-thick slide body. The Novochernomorian deposits consist mainly of pelagic coccolith-sapropelic muds with an admixture of silt and fine shell detritus. Stations 12 and 16 revealed more complete sections. At station 16, the Novoeuxinian includes pebbly mudflow deposits with a high content of shell and plant detritus. The Drevnechernomorian deposits comprise an alternation of pelagic sapropelic muds and fine, sometimes graded, layers of silt, coquinas, and autochthonous (sapropelic) distal turbidites. The Novochernomorian deposits consist of pelagic muds with shelly and silty (turbidite) interlayers and of sediments (0.5 to 2 cm thick) of pebble-shelly debris flows. The sapropelic muds contain abundant numerous plant fragments, increasing in size up the layer. There are members of lens-like shaped clay muds and brown sapropelic and silt layers. In the core from station 12, the Novoeuxinian sediments correspond to an alternation of 10- to 20 cm-thick pelagic muds and beds of shell detritus and gravel embedded into a muddy matrix, sometimes poorly graded, whose thickness ranges from 5 to 10 cm (debris flow deposits). Above, on the eroded surface, lie the Drevnechernomorian pelagic sediments of autochthonous distal turbidite type and then, Novochernomorian flat-bedded coccolith and sapropelic muds with an admixture of silty material.

In the lower part of the continental slope, the thalweg deposits are substantially different in both studied structural systems. In the eastern one, the base of the section (station 41) is composed of the Novoeuxinian muds with thin silt interlayers and poorly pronounced traces of sliding. Higher, with an erosional contact, lie poorly sorted deposits of shell, gravel and pebble, cemented by a clayey matrix (thickness 2 - 3 cm). There are also sapropelic muds with an admixture of dispersed silt and shell detritus, which contain lens-like millimeter-thick layers of pure clay and interlayers of black, coarse-grained silt similar in thickness. In its origin, the sediments of this member are related to debris flow, pelagic, and distal turbidite accumulation. Clay-free silt interlayers can be interpreted as deposits of grain flows formed on the canyon's banks. Lenses of light mud were apparently formed as a part of autochthonous distal turbidites originating from the initially suspended clayey sediments. Some sapropelic mud layers have undulated upper boundaries, an apparent indication of bottom currents. In the upper part of the core, the lens-like aggregations of coccoliths appear and the content of plant debris increases, whereas that of shell detritus decreases.

In the western canyon system, we studied thalwegs of the same depths at stations 8 and 10. Station 8 is at a junction of several canyons. The core from this station is characterized by such features as an absence of the Drevnechernomorian sediments and condensed thickness of the Novochernomorian and Novoeuxinian deposits. The base of the section is composed of the Karangat strata, including coccolith and sapropelic muds with often lenticular or discrete thin silt interlayers. In the sapropelic mud, a constant admixture of fine silty material with a predominance of mica flakes is observed. There are several interlayers (1.5- to 2-cm thick) of silts with graded bedding and a single layer (14-cm thick) composed of sand, grading upwards into silt and then into sapropelic mud with small silt lenses. This bed is most likely a turbidite member, representing the "a–b–c" elements of the Bouma sequence. Above, the erosional boundary is overlain by silt with a very low (15%) clay content (grain flow). In the upper part of the Karangat section, turbidite interlayers disappear and a flat-lying alternation of light mud, sapropelic, and coccolith layers (pelagic sediments) is observed. In the base of the Novoeuxinian section, there is a poorly graded layer (5-cm thick) of silt, gradually progressing upward into a pure bluish-gray mud. The Novochernomorian deposits are represented by coccolith and sapropelic flat-bedded muds with an insignificant admixture of dispersed fine silt.

At station 10, a fairly thick member of the Novoeuxinian deposits, overlain by a Holocene slide body, is recovered. The lower 43 cm of the Novoeuxinian sequence is represented by gravel and pebble deposits of the debris flow. The pebbles are orientated irregularly in the lower part of the bed, across the bedding in the middle part, and along the bedding at the top. Higher, above a distinct erosional surface, lies fine-grained sand with shell detritus. The sand has a gradual transition to silt and, still higher, to a member of alternating clayey and silty layers, with a progressively decreasing upward number of the latter (turbidite). The thickness of the graded bed is 11 cm. Higher, up to the top of the Novoeuxinian section, there is an interbedding of bluish-gray muds and a thin (2 - 3 cm), flat-bedded series of alternating silty and clayey layers. This member contains several interbeds of coarse-grained silt with shell detritus. The characteristic features of these sediments are a lack of clay, sharp layer contacts with the lower sometimes being an erosional one, and thicknesses varying from 1 to 3 cm (deposits of grain flows). The Holocene part of the section shows a lense-like interbedding of bluish-gray clayey mud and greenish-brown sapropelic sediments. This member is characterized by slumping folds, flow stains of variegated mud, and sometimes by lenticular, often curved, silt layers.

Submarine fans. In the eastern canyon system, channel facies were recovered at station 44. Here, the Karangat deposits are represented by pelagic coccolith and sapropelic muds with flat-bedded, and sometimes lense-like, silt interlayers (up to 0.5 cm in thickness) decreasing in number upward. A single silt layer with a distinct unidirectional cross-bedding is observed. The Novoeuxinian deposits consist of a slide member and then of a lenticular alternation of light and brown muds with curved silt lenses. Slumping folds are rather distinct. The Drevnechernomorian deposits are represented by a member of microlaminated sapropelic mud. A base of the Novochernomorian deposits contains coccolith and sapropelic mud with lenses of silt and bluish-gray Novoeuxinian muds, which have slumping structures. There are silt lenses with dispersed clay pebbles. Above this slide member, there are pelagic deposits with millimeter-thick interlayers of silt with graded bedding according to smear slide data. Apparently, this member should be interpreted as a combination of pelagic sediments and distal turbidites.

In the western canyon system, channel facies of station 4 are represented, within the Novoeuxinian part of the section, by spotted muds with rare thin layers and lenses of silt, deformed by distinct slumping folds. In the lower part of this member there is an interlayer of silt (2 cm in thickness) with a graded bedding and erosional lower contact. The Drevnechernomorian deposits include gray mud, enriched in layers of sapropelic and silty material (pelagic sediments and distal turbidites). The Novochernomorian section represents an interbedding of flat-lying beds of clay, nannofossil ooze, and sapropels with whitish layers of diatomites.

The interchannel facies of submarine cones were discovered at stations 1 - 3 and at station 45. In station 3, only the Novochernomorian deposits are penetrated. They include an alternation of clay, sapropel, and nannofossil ooze (pelagic sediments). The maximum (up to 40%) content of silt is confined to the sapropelic layers. Silt occurs in isolated layers (up to 5 cm in thickness) with distinct and often uneven contacts.

There is a common filling of a layer contact relief with silty material. Some silt beds are enriched in large (up to 1 cm) fragments of wood. They have sharply defined bottom surfaces and a gradual top boundary owing to an upward decreasing amount of silty material and plant debris. Evidently, these beds are turbidites.

The core from station 2 is mainly represented by pelagic sediments. They contain thin silty layers and thicker (up to 5 cm) layers of sand with abundant shell and plant detritus and distinct boundaries. Such a structure, together with a low content of clayey material (15%), indicates them as deposits of grain flows and turbidites accumulated over the interchannel areas. A similar section is discovered at station 1. We interpret thin, lense-like silty layers, sometimes having undulated surfaces, as deposits of bottom currents. Cyclic alternation of silty, sapropelic, nannofossil, and clayey layers, gradually changing into each other, is typical of the sequence from station 45. The cyclites formation is related to a combination of pelagic and turbidity sedimentation. Silt layers within pelagic deposits from this core are characterized by ripple marks, lenticular, sinuous, and cross bedding (deposits of bottom currents).

Banks of canyons. Canyons' walls have been sampled by five cores (stations 26, 27, 29, 31, and 33) in the eastern area and by four cores (stations 14, 19, 20, and 23) in the western one. All slope deposits are distinct by the frequent occurrence of landslide structures. Many cores display an erosional elimination of one or several stratigraphic intervals out of the section. In rare cases, we came across muds with pebbles. Isolated silty layers have a cross-bedding. In cores from stations 19 and 23, the slide deposits of Novoeuxinian and Drevnechernomorian age are overlain by thin Novochernomorian pelagic sediments.

Submarine elevations and intercanyon areas. All elevated parts of the continental slope are covered mainly with pelagic sediments, which have different compositions at various levels. On isolated elevations, they comprise almost the entire section of the Holocene (stations 5 and 37). Intercanyon areas display minor structural features of slumping, lense-like cross-bedded silt layers, and, sometimes, indications of erosion. These features are most distinctly expressed within the eastern area in the Novoeuxinian and Drevnechernomorian sediments. Smear slide studies show that silt layers have graded bedding. The presence of these strata is evidently related to the pulses of autokinetic flows.

CONCLUSIONS

Our study revealed a large variety of sedimentary processes within the studied regions of the Black Sea. Genetic types of sediments, formed by these processes, replace each other regularly both along the canyon systems and across them, reflecting different depositional environments. Whereas the variability of sediments along the canyons is controlled by various types of autokinetic flows and by the succession of suspended material deposition, the mechanism of sedimentation in transitional zones from watershed (intercanyon) areas to canyons and fans is drastically different.

Each microform of relief, such as thalwegs, banks, and watersheds of canyons, other elevations, and structural elements of submarine fans has its own set of sedimentary genetic types. This makes possible a reliable identification of the above mentioned relief forms, based on the section lithology.

A comparative analysis of lithological profiles across the strike of the canyon systems shows that the upper, middle, and lower parts of the continental slope differ primarily in the type of sediments filling the canyon thalwegs and the bottom morphology. To a lesser extent, it refers to the sedimentary series formed on the other types of microrelief. The most diverse sediments of gravitational flows have their origin in the middle parts of the canyons. Here, in thalwegs, along with deposits of debris, clastic, and grain flows, there are widespread turbidites with a complete set of the Bouma sequence elements, whereas in the lower parts of the canyons, the turbidites have an incomplete Bouma sequence.

The sedimentary patterns of all deposits are grossly dependent on erosional processes. The latter destroy the sediments in thalwegs and create numerous slide bodies on canyon's banks. They also contribute to the formation of sedimentary masses composed by local basin muds to be later redeposited by gravitational processes in the lower parts of the canyons and submarine cones. Slide bodies are typical for the whole canyon system of the continental slope. They consist of the local material only and are confined to the floors and banks of the canyons.

Sedimentary members similar in features to the distal turbidites include the distal turbidites proper that were deposited by weakened suspended flows in the lower part of the canyons within the submarine cones. They also include the sediments redeposited by short-term gravitational flows of low energy. These flows transport local intrabasin material from the canyon banks and intercanyon areas or products of canyon floor erosion. Finally, the group includes the sediments of high energy flows splashed onto the banks of canyons and watersheds. The latter type of sediments may be denoted as "subdistal turbidites."

They are distinct in the presence of lense-like interlayers of fine silty terrigenous material within the strata of pelagic muds. These interlayers display graded and cross-bedding in smear slides.

There is a certain correlation between oscillations of the sea level and the intensity of the slope processes. Moreover, canyons were active not only during a low position of the sea level (in the Novoeuxinian), which is characteristic for the western locality, now cut off the external sedimentary source areas. Activity of autokinetic processes took place in the Holocene as well. This is reflected in the erosion of relevant deposits within the upper and middle parts of the canyons

and in the subsequent redeposition of this material onto their lower parts and in submarine cones. At this time, the Shakhe River canyon, directly connected to the river, was the most active. This is displayed in debris and clastic flow traces in the upper parts of the canyon, in the erosion of bank deposits to depths of 1000 to 1300 meters, and in the presence of the Holocene flow sediments in the lower part of the canyon and in the submarine fan.

The discovery of the Karangat and Chauda deposits in different parts of the canyons and submarine cones makes it possible to trace a canyon of regime back to the Early Quaternary time. The Karangat deposits are represented by a combination of pelagic muds and turbidites formed during a high sea level. On the one hand, it indicates a canyon activity that was similar to the Holocene one. On the other hand, it is evidence of a better preservation of older deposits, judging by their thickness, in comparison to sediments of the Holocene. The discovery of the Chauda flow deposits in the canyon's thalweg at a depth of 1500 m is the first evidence of the existence of canyons in at least, the Early Quaternary time, and perhaps even since the late Pliocene.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 56 - 68.
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Evolution of the biosphere

Depositional Environments of the Lower Formations of the Riphean Sequence, Northern Part of the Turukhansk Uplift, Siberia

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Manuscript received July 12, 1992

Abstract – The author provides the facies characteristics of the silicoclastic-carbonate deposits in the lower horizons of the Riphean section of the northern Turukhansk uplift. Based on their textural analysis, the Bezymyannyi, Linok and Sukhaya Tunguska formations are considered in terms of their depositional environments. The bulk of the sediments are shown to have accumulated in settings that were much more hydrodynamically active and bathypelagic than was suggested before.

Recently, scientists have revealed that ecological settings and depositional environments strongly impact the taxonomic composition of stromatolite and microfossil assemblages, the fossil groups widely used in Precambrian biostratigraphy. For this reason, to be able to correctly estimate the biostratigraphic possibilities and the stratigraphic resolution of microfossils and especially of stromatolites that are organosedimentary structures, one must reconstruct in detail the conditions under which the fossiliferous strata were formed.

In the current Russian literature, unlike in the English, the Riphean depositional environments are considered very generally, if ever at all (e.g., Maslov, 1988, 1989), and are often treated quite contradictorily (compare Serebryakov, 1975; Krylov and Zavarzin, 1988). Essentially, after S.N. Serebryakov's contributions (1975), nobody examined the ecological aspects of the Northern Eurasia Riphean again. The same is true about the Riphean section of the Turukhansk region, which is used as a reference to substantiate the stratigraphic scale of the Northern Eurasia Upper Precambrian. The available publications analyze only some aspects of its stromatolite-forming environments (Serebryakov, 1975). Connecting the sections and certain geochemical parameters, some authors examine the general sedimentary settings of the basal silicoclastic formation of the Turukhansk region (Ivanovskaya and Afanas'iev, 1970; Bazhenova *et al.*, 1970) and touch upon the genesis of its deposits (Kozlov *et al.*, 1973; Ivanovskaya *et al.*, 1982). With the view of bridging the above gap, in 1989 - 1990 I investigated the lower horizons of the Turukhansk Riphean section in its northern, least studied part, the basin of the Bol'shaya Shorikha River (Fig. 1).

The regional stratigraphic scheme of the Turukhansk region was elaborated in the 1940s. At the moment, it does not bring about any serious discords, except contradictions concerning the stratigraphic position of the Rechka and Durnoy Mys formations, which we will not consider here. However, the opinions on the position of the Middle-Upper Riphean boundary in the section under study are controversial (e.g., Semikhatov and Serebryakov, 1983; Komar, 1990). I support the opinion expressed by M.A. Semikhatov and S.N. Serebryakov (1983), who place this boundary at the top of the Sukhaya Tunguska Formation.

The northern part of the Turukhansk region exposes a thick (over 4 km) sequence of Riphean silicoclastic-carbonate deposits. The sequence builds up two gentle westward-dipping monoclines. They are separated by a regional submeridianal thrust-fault and are broken by a series of transversal small strike slip faults. Likewise, throughout the Turukhansk region, the base of the Riphean deposits is not exposed. The deposits are discomformably overlain by subhorizontal Vendian-Cambrian beds. The section is subdivided into eight formations. The three lower ones make up an independent stratigraphic sequence, which is bordered at the top by an erosion surface. These formations are the following: the silicoclastic Bezymyannyi Formation (900 m thick), the Linok Formation of mudstone and limestone (270 m thick), and the dolomite Sukhaya Tunguska Formation (680 m). The formations are inter-

related by gradual transitions. The depositional environments of these formations will be described below.

THE BEZYMYANNYI FORMATION

Observations. The formation is a sequence of mudstones and subarcose sandstones with an apparent thickness of 900 m; it is divided into three subformations (Fig. 2). The middle subformation clearly demonstrates a cyclicity that is not found in the lower subformation and is poorly expressed in the upper one.

The idealized cycle of the middle subformation actually demonstrates the entire set of lithotypes observed in the section. It reveals a gradual upward transition from mudstones to coarse-grained sandstones. The cycle contains three units (Fig. 2).

The lower unit is composed of dark-grey and black mudstones, siltstones and sandstones; the silicoclastic material is obviously becoming more coarse farther up the section. The base of the cycle is made up of pure mudstones with thin, even flat bedding; somewhat higher, they contain a silt admixture whose content grows upsection, giving rise to interbeds of dark-grey, occasionally greenish-grey mudstones. Appearing next are lenses and lens-shaped intercalations of fine-grained sandstones, which increase upward in thickness from a few centimeters to 10 - 20 centimeters. Simultaneously, there also appear signs of erosion, low-angle cross-bedding in sandstones, and sometimes graded bedding in silty sandstones. In some cases, the top of sandstone beds displays almost symmetrical ripple marks. The upper part of the first unit is characterized by the structure of gutter casts (Whitacker, 1973) penetrating into the underlying mudstone-siltstone beds to a depth of 5 - 10 cm and filled with fine-grained sandstone and mudstone. The most frequent here are isolated gutter casts (Fig. 2a, 2b); paired ones are less frequent, and likewise, isolated tubular bodies. The walls of the gutter casts are smooth or longitudinally striated; they are occasionally complicated by narrow, tapered vertical scours (Fig. 2b). The shape of the gutter casts testifies to the erosion of well lithified sediments; the absence of grading within them and their thin, subhorizontal lamination shows that the gutters were filled under relatively quiet hydrodynamic conditions. The lamination in the cross section is also even and flat, with the characteristic onlap over gutter walls. In some cases, the upper part of the first cycle unit shows solitary interbeds of coarse-grained sandstones and fine gritstones with symmetric ripple marks on the top. The thickness of this unit varies from several tens of centimeters to several meters.

The second unit of the cycle is composed of fine-grained dark grey sandstones with rare interbeds of siltstones and mudstones, their quantities and thickness decreasing upward. The sandstones contain quartz (70 - 80%), plagioclase (oligoclase-andesine, 10 - 20%), orthoclase (1 - 5%), muscovite flakes confined to the fine-grained fractions, and rock fragments (microquartzites, 1 - 5%). The cement is quartzy, regenerated,

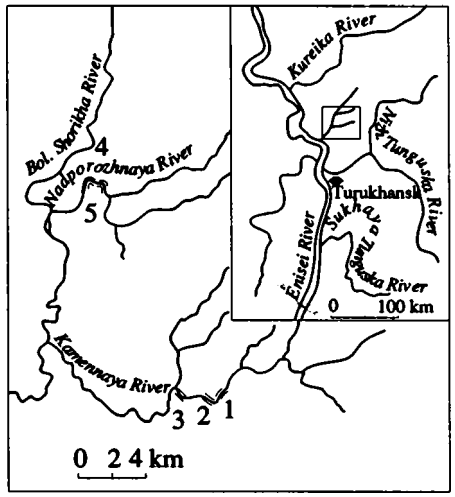

Fig. 1. *The studied region and outcrop locations. 1 – the Bezymyannyi Formation; 2 – the Linok Formation; 3 – the Sukhaya Tunguska Formation; 4 – the upper subformation of the Linok Formation; 5 – the Sukhaya Tunguska Formation.*

or argillaceous (illite and chlorite). The sandstones almost always demonstrate thin layer-to-layer granulometric sorting. There are rather rare thin beds and lenses of coarse-grained sandstones and gritstones. The sandstones are platy, and their plate thickness varies widely from 1 - 5 to 10 - 30 cm, increasing toward the top. The sandstones are characterized by fine, flat lamination, and, less frequently, cross wave-like bedding; occasionally, there is hummocky cross stratification (Dott, Bourgeois, 1982). The sandstone beds are almost always separated by thin (up to 3 mm) films of black mudstone with large quantities of fine micaceous detritus. The bedding surfaces show abundant structures of the Rhyson- etron type (Hofmann, 1971), which in a plane view look like polygonal systems similar to dessication cracks; they also have certain elements resembling worm tracks (Fig. 2c). In terms of their volume, these are tubular canalicules, 1 - 3 mm in diameter, filled with sand material; they seem to be wrapped in a carbonaceous-clayey film, tenths of a millimeter in thickness, and pressed into the overlying and underlying beds of sandstone.

The polygonal pattern on the bedding surfaces suggests that the canalicules were formed along shrinkage cracks: their rounded shape resulted from the later deep volume transformation because of the difference in the shrinkage coefficients of the sand filling the cracks and enclosing clays. During such a process, an initially wedge-shaped form, or any other form, polygonal in its cross section, will strive to become spheroidal. The plane-polygonal, incomplete systems of cracks could be formed both in subaerial environments, when dessication did not involve the entire sediment, and in subaqueous settings at the water-sediment interface (Plummer, Gostin, 1981). It is also possible to suggest that the cracks evolved during the early diagenesis

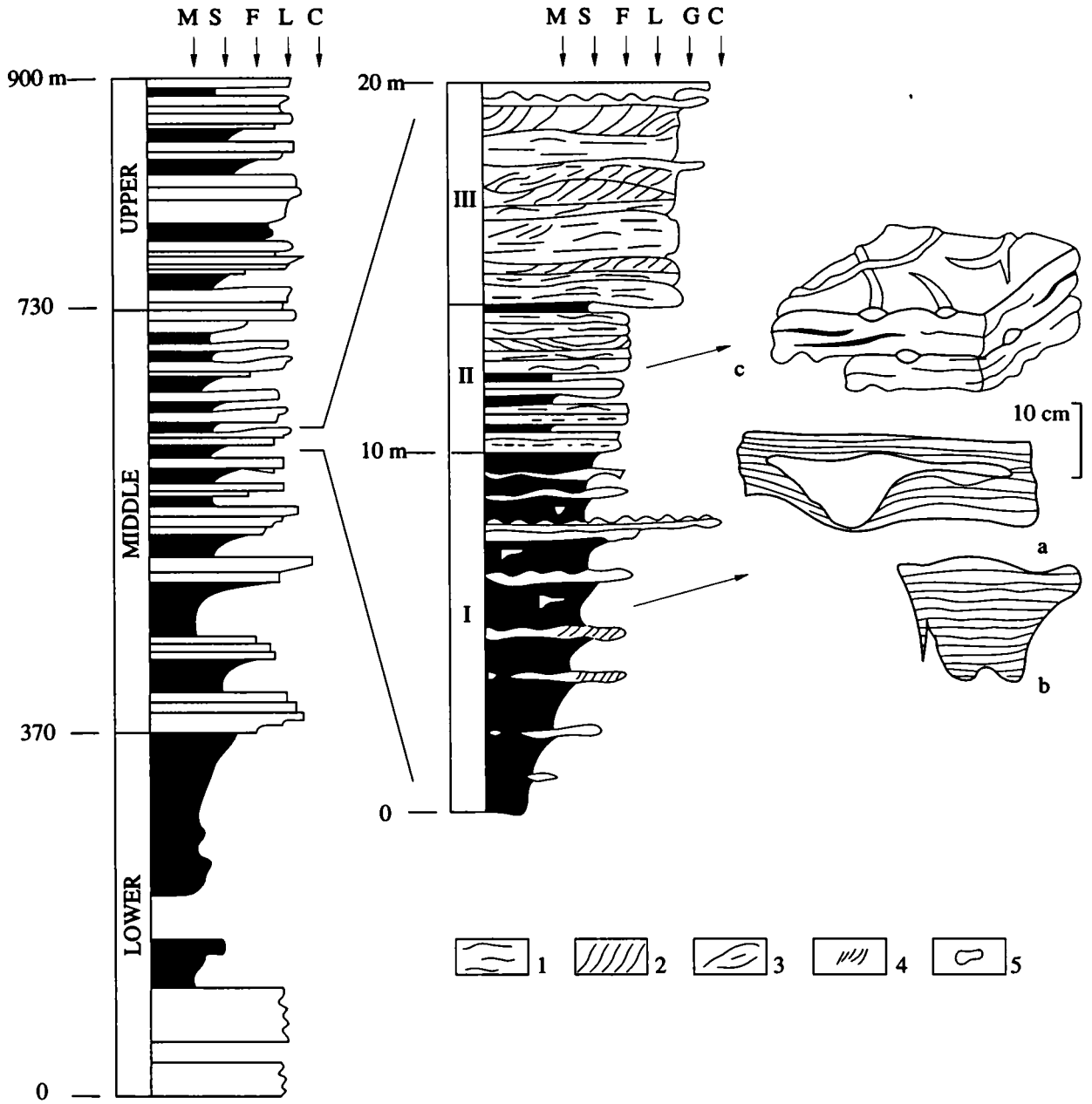

Fig. 2. *Sections of the Bezymyannyi Formation along the Kamennaya River and its sedimentary cycles. The vertical arrows indicate the rock composition: M – mudstone; S – siltstone; F – fine-grained sandstones; L – medium-grained sandstones; C – coarse-grained sandstones; G – gritstone; 1 – coarse flat bedding; 2 – large-scale cross bedding; 3 – hummocky cross stratification; 4 – small-scale cross bedding; 5 – gutter cast structures. I, II, III – the cycle units; a - c – sketches of the structures (see text for description).*

(syneresis), more so because of their confinement to the thinnest clayey strata (Kidder, 1990).

The third unit of the cycle is built by thick-platy, light-grey, well washed and sorted, medium- and coarse-grained sandstones. They contain grains of monocrystalline quartz (60 - 70%), plagioclase (oligoclase-andesine, 10 - 30%), and rare rock fragments (not more than 10%). The cement consists of regenerated quartz, usually with the completely masked original shape of the grains. It is only because of the outlines of unregenerated grains of feldspars that we can judge that they were well rounded. The sandstone's cement also displays abundant illite booklets formed after regeneration of quartz grains. There are also much smaller amounts of chlorite. Some sandstone beds contain mottled carbonate cement, which often corrodes clastic grains. It is characteristic of these sandstones to have small inclusions of pyrite (0.01 - 0.1 mm). The structure of the sandstones is not distinct; in the handspecimen, it can usually be recognized only on weathered surfaces. The lower parts of thick beds (up to 1.5 - 2 m), as a rule, reveal coarse flat bedding; the upper parts

demonstrate cross wave-like lamination and occasional hummocky cross stratification. The coarsest fractions of the sandstones are often concentrated in the top of such beds and frequently bear ripple marks, many of which are symmetric.

As we have already noted, the cycles described above are the most profound in the middle subformation of the Bezymyannyi Formation. In the lower subformation, there is a member of light-colored sandstones at the observed section base. In terms of its composition and structure, this member can be compared with the third unit of the above described cycles, and the overlying mudstone-siltstone member, with rocks of the first unit of the cycles. The upper subformation has no distinct cycles; there are primarily light-colored sandstones, which do not differ much from those in the third unit of the cycles (Fig. 2). The Bezymyannyi Formation is terminated by a 4-m-thick bed of grey coarse-grained sandstones with diverging trough cross bedding. It has no similarity with any other member of the cycles described. The bedding in these rocks is emphasized by fine-dispersed organic matter and by segregations of fine pyrite grains. The sandstones are built from quartz (80%) and plagioclase (5 - 10%), the latter's quantity being much lower than in those composing the cycles.

Interpretation. The structural and textural features of the above cycles are characteristic of open shelf settings heavily influenced by storms (Aigner, 1985). The first units of the cycles were accumulated in such parts of the basin that were the most distal from its shore and affected by storm tails, i.e., by roiled fine sands and silts superimposed on monotonous pelitic sedimentation. Farther up the section, the number of silt-sand intercalations in the mudstones increases, reflecting the ever weakening storm events and a change in settings from distal to proximal ones, if we consider the energy zones of the paleobasin. It is quite possible that the energy zonation of the basin is directly related to both the basin depths and the distance from the shoreline. With this scheme accepted, we may suggest that in the more proximal zones, there were bottom currents, which formed gutter casts. Most likely, such zones occurred in areas of flattened relief with a morphological step at the base of which we could expect the descending gradient currents to suddenly increase their friction against the basin floor.

The second cyclic units of the Bezymyannyi Formation were deposited above the supposed step and were closely associated with the rapidly accumulating beds of fine-grained sand mobilized by the storm action and with interstorm deposition of fine sediments, whose quantities gradually decreased upsection. However, the upper strata of the second cyclic unit contain thin layers and lenses of mudstones that differ from the mudstones of the first unit by their greater homogeneity and lower amount of siltstone admixtures. Such contrasting granulometric composition may be attributed to a higher concentration of clayey suspension right after a storm, as compared with the more distal zones. These settings had another specific feature in that their sediments acquired the Rhysonetron type structures, formed either in the subaquaceous regime, or during the initial diagenesis, immediately after the sediment burial beneath new portions of storm-generated sand. It is noteworthy that the above structures are frequent in the Precambrian and are not known from homofacial Phanerozoic deposits (Hoffmann, 1971).

The coarse-grained sand layers of the third unit of the cycle were deposited in the most proximal zones of the basin. The massive structures and coarse horizontal bedding near the base of these layers may be explained by the fact that, as a result of the rapidly decreasing velocity of storm-generated suspension flows, the latter were losing their sand (Aigner and Reineck, 1982). At the same time, the upper parts of these layers, with their large-scale cross-bedded series, could be related to powerful bottom currents. It is also possible to admit that one bed corresponds to one storm event, reflecting its different phases. At culminations of storms, the hummocky type of cross stratification was formed, indicating the deepest penetration of storm waves (Dott and Bourgeois, 1982). The rare occurrence of such lamination in the Bezymyannyi Formation testifies that its rocks were deposited mainly below the storm wave base.

The beds of coarse-grained sandstone and gravel-sandstones, sometimes occurring in the first units of the cycles, and the concentration of coarse material at the top of some sandstone beds in the third unit may be considered as lag deposits (Levell, 1980). They were formed when coarse fractions, repeatedly reworked by waves and currents, lagged behind the moving mass of sediments. The fields of coarse-grained deposits are confined to shelves with active storm impacts, though they occur within a wide range of depths, from the first hundreds of meters to several meters (Leckie, 1988).

Thus, the cycle shown in Fig. 2 reflects the directed change from distal to proximal settings of the basin. Similar cycles were described from Middle Proterozoic quartz arenites of the Ortega Group (USA) (Soegaard and Eriksson, 1985). The cyclic fabric of that section was attributed to the progradation of large sand fans, caused by storm-generated currents.

THE LINOK FORMATION

The lower boundary of this formation is usually drawn at the base of the first carbonate member, preceded by a 2-m-thick layer of grey-green silty mudstones with a gradually increasing carbonate content further up the section. It is this layer that shows the transition of the Bezymyannyi sandstone-mudstone deposits to the clayey-carbonate rocks of the Linok Formation. Therefore, it is more logical and more accurate to define the base of the latter by the top of the last sandstone bed. In the described section of this formation, it is divided into three subformations (Fig. 3). Our treatment of their range differs slightly from that of our precursors (Serebryakov, 1975; Kozlov *et al.*, 1988).

Observations. The lower subformation (30 m thick) consists of light-grey clayey fine-grained limestones

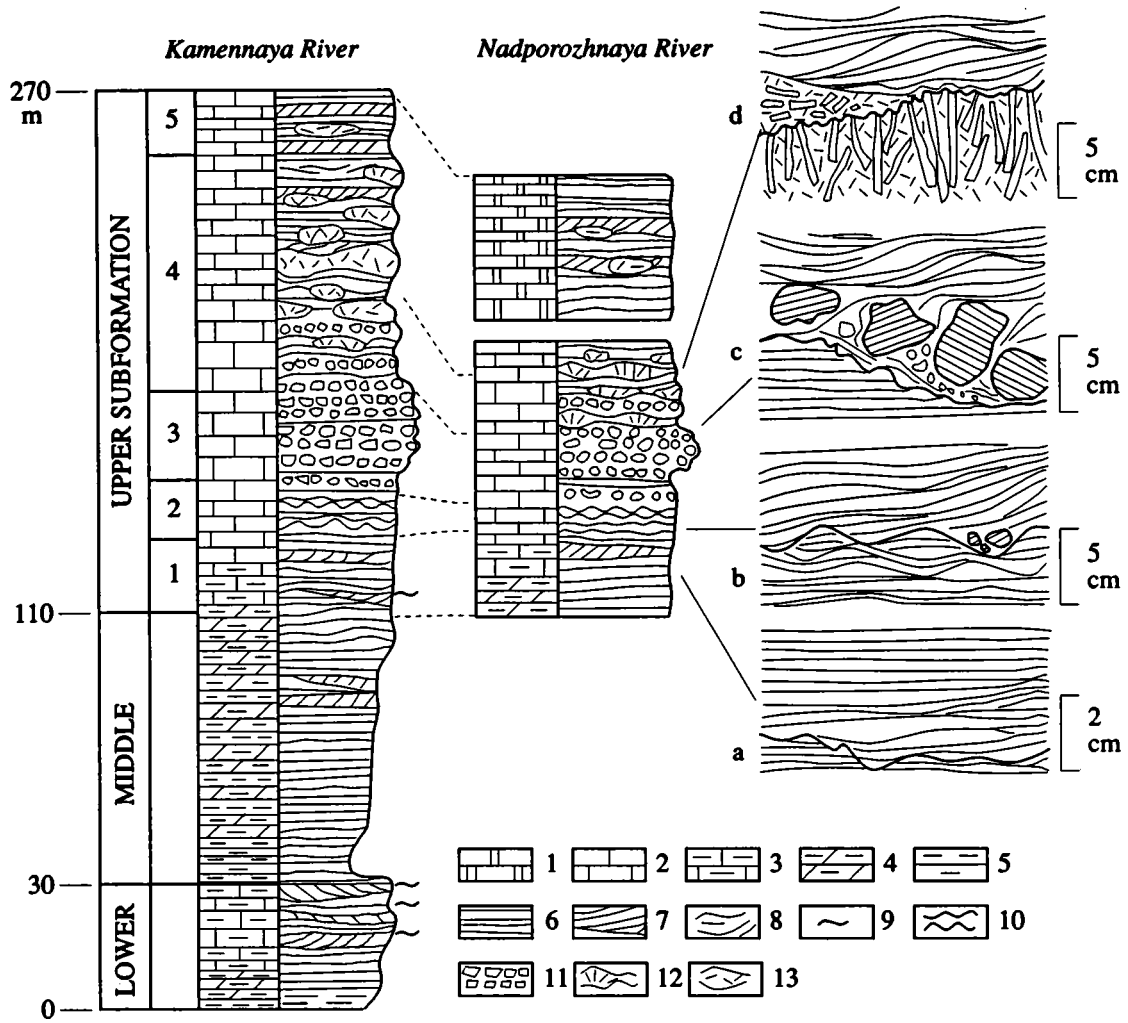

Fig. 3. *Sections of the Linok Formation along the Kamennaya and Nadporozhnaya rivers: 1 – dolomites; 2 – limestones; 3 – clayey limestones; 4 – marl and calcareous mudstones; 5 – mudstones; 6 - 9 – bedding patterns: 6 – thin parallel lamination, 7 – cross undulatory lamination, 8 – cross-wave bedding, 9 – Malginella bedding; 10 – nodular structures; 11 – intraformational breccias; 12 – lenses and lens-like beds of vertically oriented flakes and flat-pebble breccias; 13 – small lenses of subhorizontally oriented flakes and flat-pebble breccias; a - d – sketches of the structures (see text for description).*

and marls; at the base, it contains silty mudstones (1 m). The latter are massive rocks with occasionally indistinct flat bedding and mottled calcite cement appearing 10 cm above their base. Farther up the section, the calcite content grows, reaching 70% at the top. The flat lamination of the rocks is emphasized by the distribution of organic matter and clayey-siltstones, becoming more distinct farther up the subformation section. In the overlying beds of cleaner limestones (80 - 90% calcite), one finds very thin wavy lamination of the Malginella type, which was previously described as biogenic (Komar *et al.*, 1970). Thin sections show the Malginella lamination as a frequent (every 0.5 to 2 mm) alternation of fine-grained clayey limestones, containing about 80% of carbonate, and elongated lens-like layers and small lenses of 0.1 to 0.5 mm in thickness, characterized by concentrations of fine (silt) silicoclastic material cemented with a quartz-chlorite aggregate; here, the carbonate content does not exceed 30%. As seen in the outcrops, the Malginella comprises continuous beds varying in thickness from tenths to the low tens of centimeters. These beds alternate with thicker ones, consisting of flat- and cross-bedded limestones; they associate with small erosional surfaces of up to 1 cm in amplitude.

Interpretation. It seems likely that at the early sedimentation stage of the Linok Formation, there was an abrupt and rapid termination in the silicoclastics supplies, leveling off the relief and sand sediments redistribution. This is reflected in the greater maturity of sandstones occurring in the top of the Bezymyannyi Formation against the background of the rapid deepening of the basin. During the sedimentation of the lower subformation of the Linok Formation, the depth of the basin gradually decreased and the quiet-water environments gave way to those of weak bottom currents (chiefly laminar), marked by the development of normal depositional bedding, and the Malginella lamina-

tion. The latter, evidently, resulted from the activity of benthic microbial mats, which were fragmented by weak bottom currents. The mats trapped and bound the silt particles (mainly silicoclastics), emphasizing the characteristic synoptic microrelief. When there were no currents, the mats covered the sediment surface as continuous films. The remains of such films are occasionally marked by admixtures of carbonaceous material in flat-bedded limestones. With high velocities of bottom currents, the substrate was eroded and/or cross-bedded strata were deposited. Such explanation is favored by the fact that the Malginella lamination is always associated with small-scale marks of erosion and with occasional beds of cross-bedded limestones. Each level of Malginella lamination fixes a microhiatus.

Thus, the Malginella beds correspond to relatively narrow energy zones in the basin; the spectrum of the environments remained within the lower shelf and did not reach the storm wave base.

Observations. The middle subformation is 80 m thick; it is built of thin-bedded black calcareous and siliceous mudstones and marls (Fig. 3). Its lower member (8 m thick) contains black calcareous or siliceous mudstones with fine parallel lamination, emphasized by differently distributed silt and carbonaceous admixtures associated with scattered fine pyrite grains.

The upper member of the subformation is 72 m thick; it includes dark-grey calcareous mudstones, which, alongside their plane lamination, reveal individual gentle, cross-bedded layers and local scours of up to 5 cm in amplitude. The latter are sometimes marked by thin layers of siltstones and black fine-grained sandstones. The marls occasionally display the Malginella type of lamination.

The upper subformation is 160 m thick. It is composed of light-grey buff-weathering fine-grained clayey limestones. The subformation is divided into five members interrelated by gradual transitions (Fig. 3).

The lower member is 22 m thick; it consists of clayey thin-bedded limestones with thin parallel and small-scale low-angle cross or rippled lamination; there are small scours of up to 2 cm in amplitude (Fig. 3a); the lamination can occasionally be very fine, thread-like banded, and of the Malginella type.

The second member (20 m thick) shows an alternation of flat-bedded and cross-rippled limestones, similar to those in the first member, and of nodular limestones, which are thick-platy. The nodular limestones are composed of completely or almost completely isolated irregular lens-like fragments of rocks, varying from 0.5 to 5 cm in thickness and from 2 to 15 cm in length. The nodules yield either individual lenticular beds, or they chaotically overlap one another. Fine lamination is well traced within some nodules and is often truncated by the outer surface of a nodule reflecting syndepositional erosion. In other cases, it is conformable to the exterior outlines of nodules (Fig. 3b). Sometimes the beds display marks of soft-sediment deformation. The amplitude of the cross-laminated sets and scours grows syndepositional to 3 - 4 cm toward the top of the member, and its upper part shows rare fine hummocky cross stratification of up to 10 cm in amplitude.

The third member of the upper subformation (28 m thick) differs from the second by having no nodular limestones that are replaced here by intraformational breccias. These breccias alternate with subordinate layers of flat-bedded, rippled, cross-bedded, and hummocky-cross stratified limestones. The breccias are composed of isometric nonrounded fragments of thin-laminated Linok Formation limestones reaching 0.5 to 10 cm in length. The fragments are either closely grouped together or unevenly distributed within the massive limy matrix. Individual beds of the breccias do not show any size-sorting of rock fragments, though certain beds tend to concentrate larger or smaller fragments. The average dimensions of the fragments and the number of their beds obviously grow toward the member top. Most fragments bear marks of plastic deformations, yielding conformable amalgamated structures of large fragments. At the same time, some separate beds are frequently brecciated without any considerable shifts in the neighboring fragments. The top of the brecciated beds reveals fragments enveloped and capped by cross-laminated limestones, thus testifying to the synsedimentary origin of these fragments (Fig. 3c).

The fourth member (70 m) consists of limestones similar to those described from the third member, but here, they include an upward-growing amount of large, isolated lenses and beds, reaching 0.5 to 0.8 m in thickness, of vertically oriented flat fragments of fine-laminated enclosing limestones varying from 0.5 to 2 - 3 cm in thickness and from 10 to 12 cm in length (the so-called edgewise or flat-pebble conglomerates). The space between the intraclasts is filled with fine-detrital and/or pelitomorphic limestone, which occasionally demonstrates fine flat lamination perpendicular to the surfaces of the vertically oriented fragments (flakes). Beneath some of the fragments, one can see small voids filled with coarse crystalline calcite. The base of the lenses and layers of conglomerates demonstrate traces of erosion; their lateral surfaces are often steep and occasionally pendent. The tops of many lenses show the butt-ends of vertically oriented intraclasts to be overlapped by fine-laminated limestones (Fig. 3d). Very seldom, a lens may be seen as truncated by the lower surface of the overlying one. In a plane view, the interclast butt-ends provide a pattern of turbulently oriented eddies. All of the above leaves no doubt that the edgewise conglomerates are of consedimentation origin.

The amount of such lenses increases up the section and reaches a maximum at the middle of the member, then declines again. The lower part of the member still contains layers of breccias similar to those in the third member. The enclosing rocks are represented by limestones with flat, cross, cross-sinuous, and hummocky stratification.

The section of the Linok Formation terminates with a 20-m-thick member of thin-platy flat- or wave-like laminated limestones with solitary intercalations of cross-laminated varieties. This member sometimes displays thin lenses of flat-pebble breccias, in which fragments are small and occur subhorizontally.

Interpretation. The above description suggests that at the boundary between the lower and middle subformations of the Linok Formation, there was a rapid deepening of the basin, manifested by the replacement of limestones, which had different features of low-energy environments, with black siliceous or calcareous mudstones that are undoubtedly of a quiet-water environmental setting. The overlying member of dark grey marls and calcareous mudstones reflects compensated sedimentation that proceeded in the settings below the storm wave base. The entire overlying sequence of clayey-carbonate rocks from the upper subformation base up to the top of the Linok Formation may be considered as a sequence of shallowing-up basins. During its accumulation, the surface of the deposits passed through several environmental zones of the basin, which differed in their hydrodynamics. The lower member of the upper subformation was deposited in settings similar to those described for the upper part of the lower subformation. The only difference is that the amount of fine silicoclastic material in the upper subformation was extremely limited. The presence of nodular textures in the second member is associated with the generally increasing amplitudes of cross series, and hence with the activization of hydrodynamics.

The origin of the nodular structure in the limestones usually is explained either by deformation (McCrossan, 1958; Möller and Kvingan, 1988) or by concretion formation during the early diagenesis. In both cases, their sedimentary stratification may be conformable to the boundaries of the layers. However, the above-presented data suggests that the nodular structure in the Linok limestones was formed as a result of waves and currents affecting these unevenly and predominantly poorly lithified deposits. This conclusion can be supported by the fact that the nodule surfaces are enveloped by sediments with normal cross-sinuous lamination. Under such an impact of waves and currents, hardened parts of layers turned into isolated fragments, while the least lithified ones were scoured, yielding a small-hummock bottom relief or horizons of condensed nodules with a rather complex shape.

The overlying intraformational breccias of the third member can be interpreted as tempestites, and their origin may be attributed to the stratification of bottom sediments, which we shall discuss below. It should be mentioned that only in some places, the limestones are brecciated without any considerable displacement of fragments. There are frequent "one-storey" horizons of breccias where lamination within the fragments still remains horizontal. This proves that the fragments did not shift. Such horizons might have been produced by the action of the rare largest waves, which affected the basin bottom at a relatively great depth.

The flat-pebble (edgewise) conglomerates of the fourth member are morphologically distinct from the breccias of the third member. There are three opinions on the origin of these conglomerates. According to the first, they resulted from the subaerial dessication of clay-carbonate sediments in the tidal flat settings (Serebryakov, 1975; Krylov and Zavarzin, 1988). Second, they are supposed to be tempestites that resulted from the brecciation of thin limestone layers during the most intense storms (Kreisa, 1981; Sepkoski, 1982). Such processes can take place at considerable depths, below the fair-weather wave base. The third suggestion about the origin of the above conglomerates is that they are the result of the accumulation of vertically oriented pebbles ("stone rosettes") in extremely shallow settings along the shores of seas, lakes, and riversides (Dionne, 1971). The vertical position of the fragments that form "rosettes" is related to the wave action and is used to indicate the former shore lines (Ricketts and Donaldson, 1979).

The facts we have observed favor the storm origin of the flat-pebble conglomerates. The above lenses are scattered within the plane-laminated, fine-platy limestones; they do not associate with marks of dessication, local erosion, or any other inter- or supratidal features in general. Most likely, the flat-pebble (edgewise) conglomerates owe their origin to properties of the bottom sediments. When fine-platy flat-laminated limestones and interlayers (0.1 -1 mm thick) of their clayey varieties had different rates of lithification, the basin bottom became mechanically stratificated and, as a result, could become broken by wave action during the strongest storms. Between lenses or isolated intercalst layers, one finds fine-platy flat-laminated limestones deposited in quiet-water settings or cross-bedded limestones deposited in bottom currents settings. The observed layers with small-hummock cross lamination may be interpreted as tempestites produced without any breakage of the basin bottom. When these layers overlap beds and lenses of flat-pebbled conglomerates, they correspond to the final stage of a storm event.

Thus, the diversity of the structures described above might be caused by the variable mechanical properties of bottom sediments originating along with their early cementation and by the episodic storm impacts.

The above data suggests the environmental interrelations shown in Fig. 4.

The early cementation of the discussed carbonate deposits could be attributed to the oversaturation of pore water with the microbially generated hydrocarbonate in the reducing environments (Sepkoski, 1982). Pyrite concretions, rather frequently found in the limestones of these very zones, support the assumption of reducing conditions during these rocks' early diagenesis.

The facies interrelations, similar to those described above, are known from some present-day environments on the northern slope of the Bahama Bank (Mullins *et al.*, 1980), where nodular limestones occur between an abyssal zone (below 600 m) of soft carbonate muds

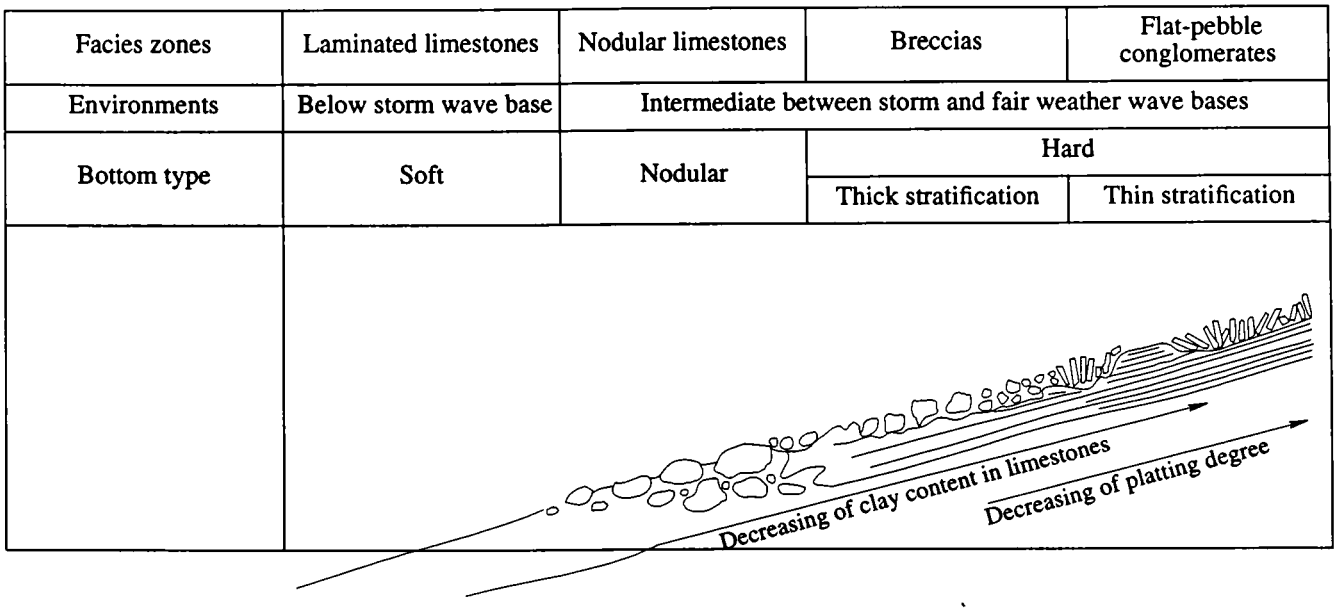

Fig. 4. *Depositional environments, bottom types, and the succession of sedimentary structures for members 1 - 4 of the upper Linok Subformation.*

and a hard-bottom zone of middle depths. The nodular limestones and lithified deposits of this region resulted from early submarine diagenesis, and the entire rock sequence follows the order of currents becoming weaker with depth. Similar facies relations were described in the Upper Cretaceous limestones of southern England (Kennedy and Garrison, 1975).

THE SUKHAYA TUNGUSKA FORMATION

Observations. Within the studied region, the formation is composed of light- and dark-grey dolomites with chert concretions in the upper part of the section and intercalations of limestones in its lower part. In the section along the Kamennaya River, limestones comprise about 7% of the total volume of this formation, but further north, in the Nadporozhnaya River section, they are not present at all (Fig. 5). The lower boundary of this formation is conventionally defined by the disappearance of the flowing type of lamination, which characterizes the Linok Formation, and by the replacement of limestones with a predominantly dolomotic unit. The Sukhaya Tunguska Formation consists of two main rock types:

I. There are dark-grey, nondetrital dolomites and limestones, which are thin- and flat-laminated, less frequently cross-laminated, with a strong butiminous smell after breaking a sample. The granoblastic texture of these dolomites is defined by crystals of different size, mainly fine or small. The limestones are also granular, but their grains are usually smaller than those of dolomites. Carbonaceous material is rather homogeneously distributed in the rocks, yielding either dark-brown rounded clots, comparable in size with crystals, or dusty inclusions inside the crystals. It is less frequently found as some film cement. Both the limestones and the dolomites reveal zonal silicification.

The overwhelming majority of these rocks have the so-called "burunduk" (banded) structure comprising a system of small syneresis cracks, filled with fine-grained grey calcite in limestones and white coarse-crystalline dolomite in dolomites. The layer cross section reveals isolated cracks, which are usually oriented normally to the bedding. Less frequently, they are inclined or horizontal, occasionally merging together and producing an intricate reticular pattern. The cracks vary in width from 1 to 10 mm; their borders are sharp and even. The cracks deform the fine lamination of rocks, yielding polygonal patterns on the bedding surfaces, which resemble those of subaerial dessication. When there are many cracks in these rocks, the latter become convolutely bedded. Banded structures are of syneresis origin, i.e., they result from dehydration of the sediments and the reduction of their volume during diagenesis. The deformed lamination appears due to different coefficients of post-syneresis compaction of sediments and coarse crystalline carbonate filling the cracks.

II. Along with the dark-colored varieties, the Sukhaya Tunguska Formation contains grey to light-grey dolomites and limestones with detrital textures of wackstone and packstone. Two types of detritus can be recognized in these rocks. The small grains (0.1 to 2 mm) are spheroidal (pseudo-oolith) or irregularly oval. They consist of micrites and are set into the fine-grained carbonate matrix of the granoblastic texture, often being rimmed by a zone of mosaic coarse-grained carbonate and, more seldom, by quartz. In some cases, the fragments have dark-brown and distinct rims of about 0.05 mm in thickness. The fragments are not sorted. As a rule, the dolomite varieties of lightest color

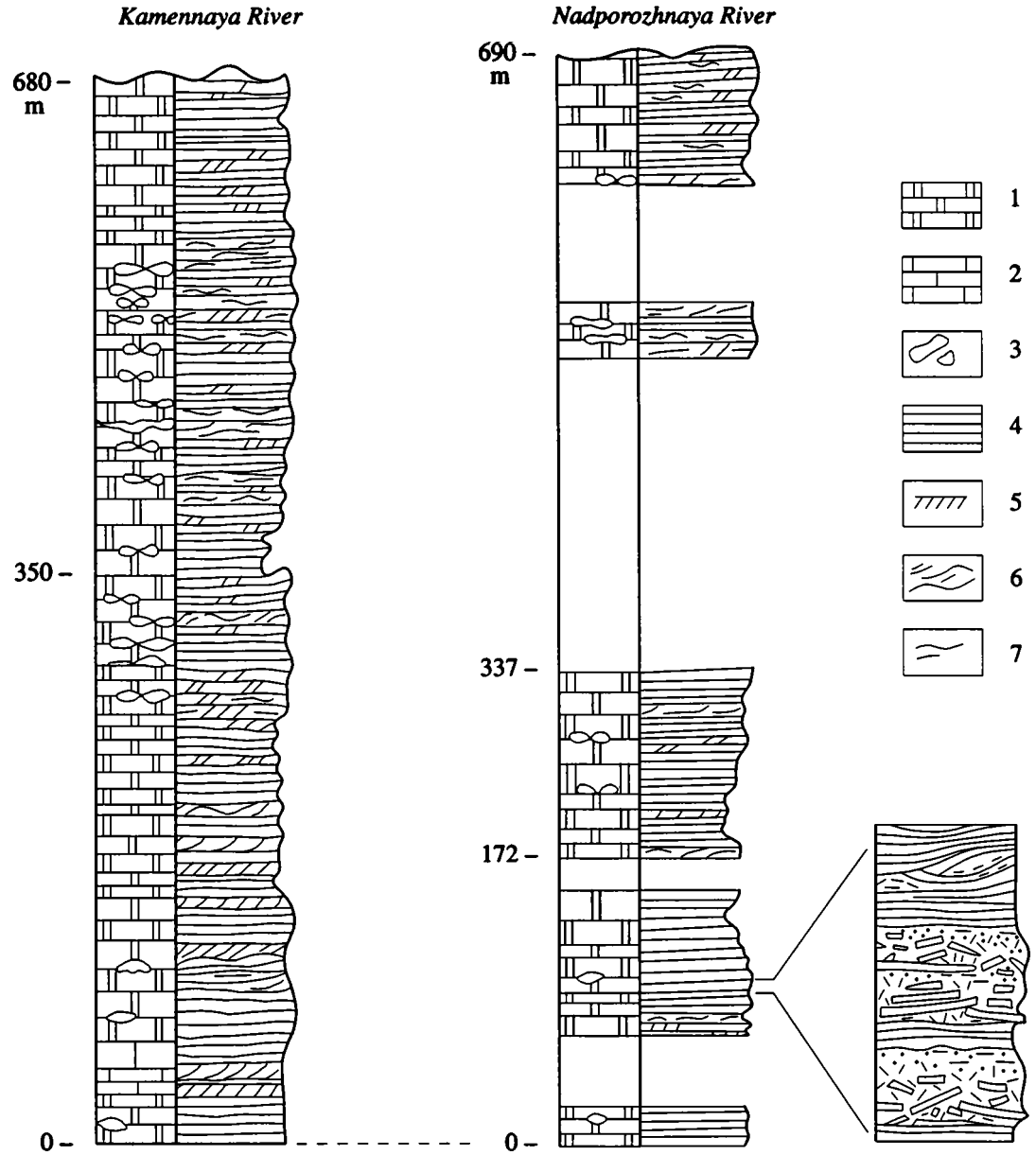

Fig. 5. *Sections of the Sukhaya Tunguska Formation along the Kamennaya and the Nadporozhnaya rivers: 1 – dolomite; 2 – limestone; 3 – chert concretions; 4 - 7 – patterns of bedding: 4 – fine flat bedding, 5 – cross bedding, 6 – large-scale cross-sinuous bedding, 7 – small-scale flat-sinuous bedding.*

prove to be the most coarse-grained and undergo a partial or complete recrystallization of detrital micrite, yielding shadow pseudo-ooidal textures.

Large fragments, up to 20 cm long and 2 cm thick, consist of fine-grained dolomite and represent debris from the enclosing beds. In some cases, they are oriented along cross-bedding, but they usually lie subhorizontally. Most frequently, the fragments are not at all rounded. Sometimes they are semi-rounded and flat. On the whole, the smaller the fragments, the greater their roundness. Occasionally, some large fragments reveal signs of soft-sediment deformation, such as intraclasts of broken beds underthrusting one another and other marks of deformation in the unconsolidated state.

Detrital carbonates are characterized by flat, cross-sinuous lamination and hummocky cross stratification; many beds have coarse graded bedding with coarse-grained packstones at the base and medium-grained packstones and wackstones at the top.

Some levels of the Sukhaya Tunguska Formation display members (3 to 10 m thick) of fine cyclicity. The lower boundary of such cycles is sharp, occasionally erosional. Their lower part consists of graded packstone, and the upper is represented by flat-bedded, fine-grained wackstones (Fig. 5). The base of the cycles sometimes may have marks of sediments' breakage and *in situ* burial and also silicified surfaces (hard ground) with signs of erosion. Individual cycles reach the low

tens of centimeters in thickness. The cyclic parts of the section are separated by thickness-commensurable piles of fine-grained wackstones with horizontal or cross lamination. Sometimes one finds here beds of fine-platy and fine- and medium-grained massive dolomites, often with flat or gentle ripple bedding, which include chert nodules. The number of such beds drastically increases in the upper part of the Sukhaya Tunguska Formation.

Interpretation. Cyclic sedimentary sequences with distinctive graded bedding and occasionally hummocky cross lamination, which resemble the Sukhaya Tunguska deposits, have been described as carbonate tempestites (Kreisa, 1981; Aigner, 1985). In an idealized cycle of such a tempestite, the leading element is represented by the bed couple, composed of graded detrital and micritic carbonate. The first member of the couple corresponds to episodes of storm sedimentation, the second, to those of interstorm deposition. However, in our case, the place of micritic carbonate in the described cycle was occupied by a fine-grained carbonate.

There are some reasons for considering dark carbonates with thin, flat lamination as deposited in quiet-water settings of the deepest basin parts located below the storm wave base. These carbonates progress into graded tempestites accumulated above the storm but below the fair weather wave base. Certain members of fine-platy detrital dolomites with small-scale sinuous and rippled lamination seem to have been accumulated in settings of permanent wave action; they can be considered as the shallowest-water deposits formed under the upper subtidal and, probably, the intertidal environments.

CONCLUSIONS

1. The Bezymyannyi Formation is a product of siliciclastic sedimentation in a constantly subsiding basin. The formation section, on the whole, reveals a large regressive sequence. This is reflected in a general tendency, when settings of the open, relatively deeper parts of the basin with quiet pelitic sedimentation below the storm wave base give way to those affected by the storm activity.

The cyclic structure of the section, most likely, was produced by the storm progradation of individual zones of cyclic sedimentation, followed by a rapid rearrangement of the storm current system, preventing distribution of the coarse material. The reason behind such rearrangement is still unclear. It may be attributed to changes in either the shore line configuration, or the sedimentation-feeding flows on land. The studied cycles can be compared with similar Upper Precambrian and Lower Cambrian cycles from North America (Soegaard and Eriksson, 1985; Simpson and Eriksson, 1990), described in terms of a three-dimensional facies model. The comparison permits concluding that in the Bol'shaya Shorikha River basin, we have rather distal facies.

2. At the boundary between the Bezymyannyi and Linok formations, the sedimentation environments underwent cardinal changes that were probably initiated by the onset of a large transgression and general deepening of the basin. The next most essential deepening of the Linok basin is marked at the boundary between its lower and middle subformations whose sediments, in general, were accumulated below the wave base, in quiet deep-water environments affected only by weak bottom currents. Deposits of the upper subformation display marks of episodic storm events, which become more frequent upsection. Such marks testify to a gradual decrease in the basin depth and to an increase of the storm wave action. The very fact of such action can, at least, prove the environmental conditions of the lower supertidal zone.

3. It seems that at the beginning of the Sukhaya Tunguska accumulation, the basin deepened again. The sedimentation proceeded generally in quiet-water settings below the storm wave base. In due time, the depositional environments became shallower, affected by some storm impacts and occasionally by fair weather waves. The upper, chert-bearing part of the Sukhaya Tunguska section corresponds to the shallow-water settings. Some of these beds were deposited in the environments of the upper supertidal and, most likely, the intertidal setting.

4. Generalizing conclusions based on this study, one can state that the three lower formations of the Turukhansk uplift were deposited in such settings whose depths and hydrodynamic activity were much greater than was suggested before. In terms of their lithology, the discussed deposits of the Turukhansk region are extremely similar to those of the Kerpyl' Group of the Uchur-Maya region and the upper part of the Sukhoi Pit Group of the Irkineevo inlier in the Enisei Ridge (Semikhatov and Serebryakov, 1983). Taking this similarity into account, one may consider the above conclusion as having not only regional, but much wider importance.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 69 - 79.
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The Paleozoic Boreal Biogeographic Region

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Manuscript received August 10, 1992

Abstract – The main features of the Boreal biogeographic region in the Middle and Late Paleozoic are characterized and their similarity in principle with those of the present-day Boreal region is shown. The region arose in the Silurian; it comprised the Siberian craton southern margin and, gradually growing, by the end of the Paleozoic gained configurations close to those of the present-day. According to the paleomagnetic data, it was caused by the Middle-Late Paleozoic northward drift of the continents now situated in the Arctic.

In recent years, paleontologists' interest in paleobiogeography has been ever growing. A number of biogeographic schemes, both for the planet of Earth as a whole and for individual regions, were designed. However, essentially different methodological approaches to such schemes make it very difficult to restore the evolution of even the largest biogeographic subdivisions (biochores).

Today, the basic regularities in the distribution of organisms on the Earth are determined by climatic zonality. Accordingly, all biogeographic schemes (Zezina, 1985; Zenkevich, 1956) identify as the largest biochores the Tropical and two circumpolar regions (in some papers, they are called zones or belts), which are divided into smaller biochores. The main difference between the Tropical and circumpolar, Boreal (Subarctic), and Notal (Subantarctic) regions (belts) is the exceptional faunal diversity in the Tropical region. Thus, L.A. Zenkevich (1956) wrote that today, about 1200 species inhabit the Kara Sea, whereas in the Mediterranean Sea, their numbers increase to 7000, and in the tropical sea of Indonesia, there are 40,000 species. A number of large taxa are endemics of the Tropical region: all reef-building organisms, most plankton forms, and others.

Such regularity can be clearly traced throughout the Cenozoic, Mesozoic and Late Paleozoic. In this regard, the principle governing the identification of the major biogeographic units corresponding to the climatic zones is usually applied to the biogeographic zonation of the Cenozoic, Mesozoic and Late Paleozoic. However, it is not sustained for the Middle and Early Paleozoic biogeography.

Most authors ("Paleontologiya...," 1981; "Atlas of Paleobiogeography," 1973) consider the Devonian and Silurian Mal'vina-Kaffir province to embrace the southern circumpolar area. The Boreal region has so far never been identified by any investigators of the fauna of these periods. Thus, it was always emphasized that all Early-Middle Paleozoic deposits in the regions, occurring today in the high northern latitudes, contained typically tropical fauna (Alekseeva, 1985; Dubatolov, 1972; "Paleobiogeograficheskii Atlas...," 1979). Based on this, I (Ustritskii, 1970) also used to believe that the Boreal region arose only at the beginning of the Late Paleozoic as a result of the general cooling of the Earth's climate and reduced links between the basins of the Arctic and the Tethys.

However, according to new data, there can be another explanation as well. As paleomagnetic evidence shows (Gorodnitskii *et al.*, 1978; "Paleomagnitologiya," 1982; Ustritskii and Khramov, 1989.), during the Paleozoic, all continents, which today occur in the Arctic, used to be in the equatorial zone, while the majority of the Earth's Northern Hemisphere was occupied by the ocean. In this case, we shall have to admit that the Boreal region arose, not so much as a result of the climatic changes on the Earth, but rather, because of the Middle-Late Paleozoic northward drift of the continents. In view of this, the purpose of our paper is twofold: (1) to determine the criteria for identification of the Boreal region and to define the time of its rise and position in different periods of the Paleozoic; (2) to correlate paleobiogeographic and paleomagnetic data and to determine the degree of their mutual accordance and their significance for geodynamics.

Consideration of Paleozoic biogeography is complicated mainly by limited knowledge of the sparse fauna of the Paleozoic oceans, whose sediments, preserved and outcropping now only as fragments in folded areas, are usually strongly metamorphosed and their fauna, destroyed. As a result, zonation of Paleozoic marine basins means zonation of shelf seas, which were almost completely preserved on cratons. Such zonation is based on benthic faunal groups.

Most scholars regard planktonic and partially nektonic groups as being cosmopolitan. However, the bulk of such groups are spread only in the Tropical region and are absent in both circumpolar regions. Thus, they can be used for biogeography.

Paleobiogeography has been examined, beginning with the Silurian, period by period, or, to be more exact, by some divisions of the periods. Each period is first provided with the comparative characteristics of the regions, and then their boundaries are drawn on the current topographic maps.

The Boreal region has been characterized in greater detail. The case is that the fauna of the Tropical region is much better known to a wide circle of experts, because throughout the entire Paleozoic, the Tropical region included Western Europe, the Russian craton, and the Urals. The latter areas were studied for over 100 years, and it was this fauna that was described in all textbooks on paleontology and historical geology as the most characteristic fauna for any given period.

Regarding biogeography by periods, I focused my attention on brachiopods, which comprised the most numerous group existing in all biochores throughout the entire Paleozoic. Furthermore, I am particularly familiar with brachiopods, though I make use of all other groups of Paleozoic fauna and flora.

GENERAL CHARACTERISTICS OF PALEOZOIC BOREAL REGION

The main characteristic feature of the Boreal region, permitting its identification throughout the entire Middle-Late Paleozoic, is the fact that it lacks some major taxa. In all periods of the Paleozoic, it is devoid of colonial corals, calcareous algae, echinoids, and all reef-building organisms. There are almost no conodonts. Completely absent are Silurian graptolites, Silurian-Devonian pentamerids, Devonian *Stringocephalus*, Early Carboniferous *Gigantoproductus* and *Striatifera*, Permian Lyttoniidae and Teguliferinidae, and others. There are no Carboniferous or Permian fusulinids. Ammonoids and tetracorals can be found only sporadically. Thus, the fauna of the Boreal region essentially does not contain those basic groups that are used to substantiate the Middle and Upper Paleozoic stratigraphic scales and the zonal division of the Tropical region.

Only brachiopods, bivalves, bryozoans, crinoids, and small foraminifers are widespread within this region. However, they are considerably less diverse than in the tropics (Ustritskii, 1970). Endemic taxa, characteristic only of the Boreal region, are present in all faunal groups here, though, as a rule, their rank is low. There is one endemic family among the brachiopods and one among the ammonoids (Horridoniidae and Yakutoceratidae, respectively). At the same time, endemic genera are present even in such a group as the ammonoids, which is usually regarded as cosmopolitan. For example, in the Carboniferous, such genera as *Eoshumardites* and *Postaktubites* were endemic; in the Permian, the *Tumaroceras* and *Spirolegoceras* genera.

The number of endemic boreal genera is steadily growing as the boreal fauna became more studied. For example, ten years ago, an idea prevailed that the *Licharewia* genus, which is rather common in the Kazanian Sea of the Russian craton, was spread throughout the Arctic. R.V. Solomina (1985) made a revision and demonstrated that there are no representatives of this genus in the Arctic at all, and the forms that had been determined as *Licharewia* in fact belong to the endemic *Tumarinia, Olgerdia*, and *Penzhinella* genera.

The faunal monotony of the Boreal region is aggravated by the fact that most of endemic taxa, known within its limits, can be found only sporadically. The bulk of its fauna is represented by few genera, which are extremely widespread and frequently form monotaxonic communities (*Tuvaella* in the Silurian, *Yakutoproductus* in the Late Paleozoic, etc.).

In the Late Paleozoic, of great importance are taxa of bipolar spread. They can be found in all faunal groups. Among ammonoids, they include such well-known genera as *Paragastrioceras* and *Uraloceras*, though their distribution range is slightly wider than that of the fusulinids, whose limit is traditionally regarded as the boundary of the Tropical region. Most characteristic of both the Boreal and the Notal regions are the *Inoceramus*-like bivalves. The *Euridesma* genus is the most spread in the Notal region, and its name has been given to the relevant faunas. Though this genus was not found within the Boreal region, most of the genera (*Atomodesma, Aphanaia*, and others) are present in both regions (Astaf'eva, 1986).

Unlike fauna in the circumpolar regions, their flora differs essentially. There were numerous suggestions concerning the presence of the Gondwana elements in the Angara flora, but more careful studies did not confirm these ideas. The morphological similarity of the floras proved to be due to their convergence and not to their genetic affinity (Vakhrameev *et al.*, 1970).

As we can see, some general faunal impoverishment of the Boreal region, as compared with the Tropical region; the fact that some major taxa were absent in the Boreal region, inhabiting only the tropics; and the low rank of boreal endemics permit us to draw direct analogies between the Boreal regions of the Paleozoic and of today. Despite their differences in climate, sea level, and position of continents, they prove to be similar in principle.

While during the Middle-Late Paleozoic, the Boreal region was expanding in size; its faunal difference from the Tropical region was growing. In the Silurian-Devonian, the Boreal region, along with its characteristic endemics, still had some rare representatives of groups that were usual for the Tropical region, for example, tabulates and colonial tetracorals. Fauna in the Boreal region became most differentiated during the Late Permian, when the total number of taxa reached its minimum, and the fauna became extremely monotonous, comprising mainly endemic genera.

In the Late Permian, the Boreal region gained its maximum dimensions. During this time, it became especially heterogeneous, with the different character of its boundaries in Europe and America, on the one hand, and in East Asia, on the other.

In northern Eurasia, the fauna and flora were generally becoming more scarce when moving from the Russian craton eastward, but this impoverishment was gradual, i.e., there was a transitional zone of at least 2000 km wide (Ustritskii, 1970). Thus, the *Komia*, *Callipteris*, and *Compsopteris* genera, which were typical of the Pechora basin, had already disappeared in the eastern Taimyr; ferns were completely absent east of the Verkhoyansk area and in the entire Northeast of Russia (except in Koryak Highland). Only few cordaites were present in the zone of the ice-rafted marine sediments.

The same was true about the fauna. There were still corals (colonial tetracorals and rare tabulates) on Svalbard, but only solitary tetracorals were known east of it. Among the brachiopods, the eastern limit of the *Diktyoclostida* distribution passed through Novaya Zemlya and the Pechora basin. East of this boundary, there were almost no representatives of cosmopolitan genera, and the fauna consisted of endemics of the Boreal region (*Cancrinelloides*, *Tumarinia*, and others).

In the east of Asia there was no transitional zone, the boundary between the regions being extremely sharp. The Upper Permian deposits of the Koryak Highland, with their reef massifs and fusulinids, identical to those of the Upper Permian in the Mao Kou Formation of South China, are typical of the Tropical region. These deposits occur a mere 150 km off the Penzhina region and 400 km off the Omolon massif, a region with ice-rafted marine sediments and the poorest fauna. The width of the transitional zone drastically reduced in Mongolia and the Primor'e region. Thus, the generally regular oval of the Late Permian Boreal region (Fig. 1) proves to have no internal symmetry. Its coldest part is confined not to the center of the oval, nor to the area of today's North Pole, but to its eastern margin, the regions of Verkhoyansk and the Northeast of Russia. It is worth mentioning that it is also here that the region of the continuous distribution of ice-rafted marine deposits occurs.

POSITION OF THE BOREAL REGION IN THE PALEOZOIC

During the Silurian, all marine basins in the Northern Hemisphere were embraced by an integrated vast area, named the Cosmopolitan, or the Old World, region. It contained typically tropical fauna ("Atlas of Paleobiogeography," 1973). Various scientists tried to identify provinces within this region, always emphasizing that the differences between the identified biochores were very insignificant and that their boundaries were unsteady (Dubatolov, 1972; Ivanovskii, 1965; "Paleobiogeograficheskii Atlas...," 1979). D. Kal'o noted in 1975 that since the late Llandoverian to the end of the Silurian, an endemic *Tuvaella* community became spreading in the middle of Asia, though he believed that "the biogeographic significance of this community was not great." However, the latest paper by N.P. Kul'kov, E.V. Vladimirskaya, and N.L. Rybkina (1985) showed this fauna to be much more distinctive than was thought before. Seven endemic genera of brachiopods were found here. Some of them are of a mass spreading (*Tuvaella*, *Tuvaechonetes*, *Tuvaestrophia*). Over half of the species found here were also endemic. There were absolutely no pentamerids (except the early Llandoverian), which were rather frequent in the rest of Eurasia. Tetracorals likewise exhibited a high degree of endemism. There were practically no graptolites.

The *Tuvaella* fauna is spread within a narrow strip stretching from the Altai and Tuva through Mongolia to the Amur region (Fig. 1). Inasmuch as the environments were repeatedly changing here during the Silurian, the long existence of the endemic *Tuvaella* fauna cannot be attributed to the fact that it was confined to some definite facies, or to the isolation of the basin, as is often assumed today. We shall demonstrate that the peculiarity of the *Tuvaella* fauna results from the climatic zonality, which can be first clearly distinguished according to the character of the biota of the Northern Hemisphere in the Silurian. The *Tuvaella* fauna may be considered as a kind of progenitor of Devonian and Carboniferous boreal faunas.

During Early Devonian time, a distinctive fauna continued to develop in Mongolia, Tuva, Transbaikalia, and the Amur region. From this, one can distinctly see that the Silurian *Tuvaella* fauna was succeeded by the Devonian fauna (Alekseeva, 1988). R.E. Alekseeva (1985; 1988) noted that though the endemic genera in all groups were not numerous, the endemic species in these region0s comprised 50% of the tabulates, 80% of the tetracorals, 60% of the brachiopods, and 90% of the crinoids.

V.N. Dubatolov (1972) seems to have been the first to notice the most characteristic feature of the fauna of this region. It is not so much the high rank of its endemic corals, but rather its general impoverishment and absence of major taxa such as Alveolitidae, Heliolitidae, *Striatopora*, *Thamnopora*, *Cladopora*, and others, which was typical of the Boreal region.

During Devonian time, the Boreal region essentially expanded to include the entire Altai-Sayan region and, in the Late Devonian, the eastern part of Kazakhstan ("Paleobiogeograficheskii Atlas...," 1979). In Fig. 1, the boundary between the Boreal and Tropical regions is drawn in the same way as is presented by all experts for the end of the Middle Devonian. On the Siberian craton, this boundary is drawn between the Kazakhstan and Tungus phytogeographic regions identified by N.M. Petrosyan ("Stratigrafiya SSSR," 1973).

Beginning with the Carboniferous, the wide distribution of terrestrial flora, which is very sensitive to climatic changes, makes it possible to directly correlate paleobiogeographic and paleoclimatic data and to establish general temperature parameters characteristic of the biochores under study. In the case of the frost-free zone, conventionally called the Tropical, evergreen plants are typical, because there are no essential sea-

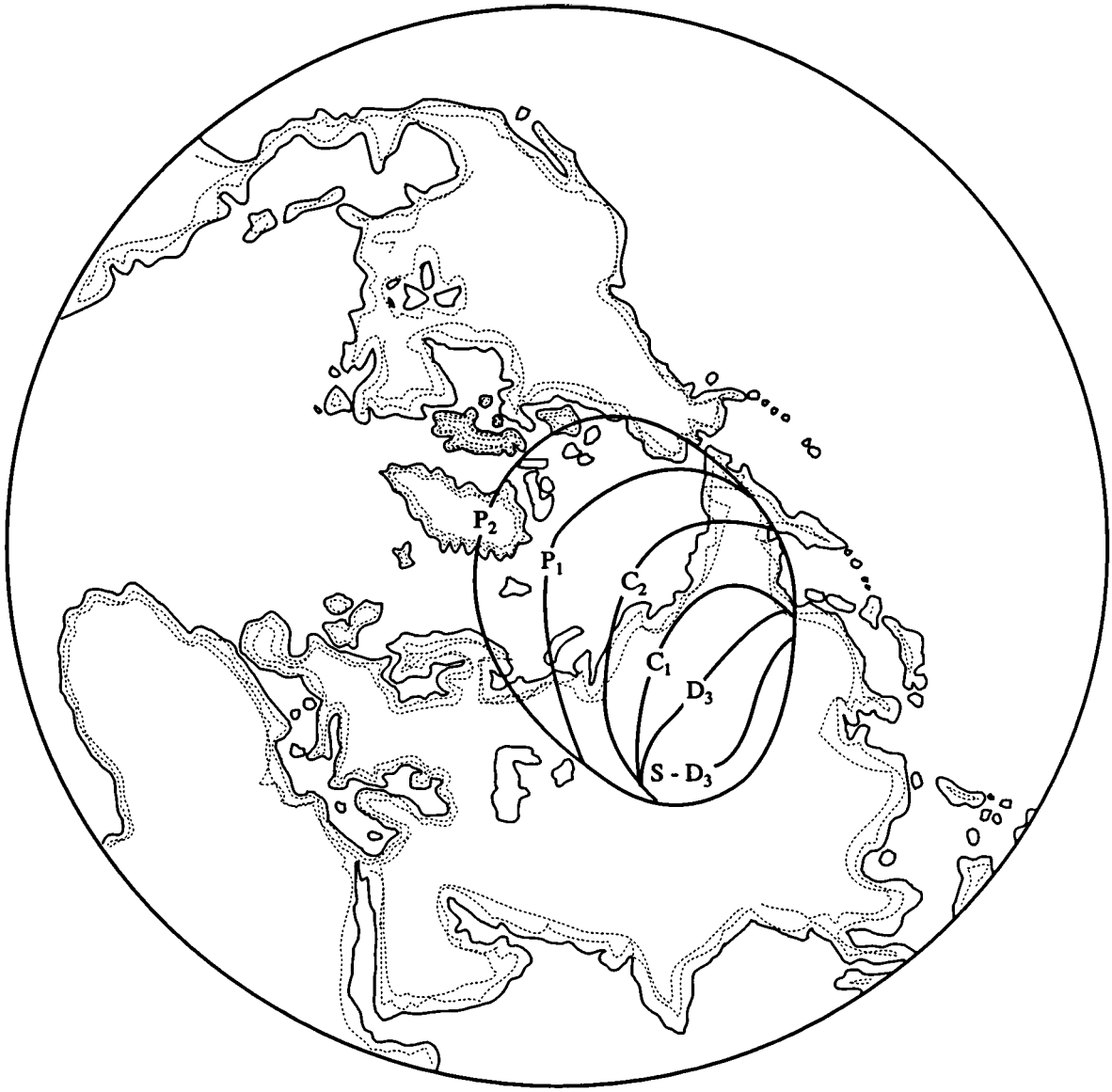

Fig. 1. *Position of the Boreal biogeographic region at different Paleozoic periods.*

sonal climatic fluctuations (no winter frosts). Such plants do not defoliate in the winter, and their wood has no annual rings of growth. The zone with steady negative winter temperatures is characterized by plants defoliating in the winter and having wood with annual rings of growth (Vakhrameev *et al.*, 1970). (The exceptions existing today, such as the recent cowberries, do not change the general picture). Accordingly, for the Carboniferous and Permian, the Angara kingdom (the Angara region in the Carboniferous) has been distinguished according to flora. This kingdom corresponded to the extratropical zone of the Earth's Northern Hemisphere. The southern boundary of the Angara kingdom coincided with the southern boundary of the zone with steady negative winter temperatures.

When analyzing the distribution of fauna and flora during the Carboniferous and Permian ("Paleontologiya...," 1981), it was found that the most considerable faunal changes, which make it possible to distinguish the Boreal and Tropical regions from each other, occurred near the boundary of the Angara kingdom, in fact coinciding with it. As a rule, in places near this boundary where one finds mixed, transitional fauna, the flora is of the same character ("Pribalkhash'e...," 1976). Thus, the identification of the boundary of the Boreal region in the Carboniferous and Permian proves to be more reliable than in the Silurian and Devonian.

All textbooks on historical geology and paleontology mention the brachiopods *Gigantoproductus* and *Striatifera*, the tetracorals *Palaeosmilia*, *Dibunophyllum*, and others, the tabulates *Syringopora*, and the chaetetids as most characteristic of the Visean. This fauna is spread from Western Europe to China and from the

Moscow region to the Northeast of Russia and was for a long time considered to be cosmopolitan.

However, in some areas of Siberia (Eastern Kazakhstan, Kuznetsk Basin, Transbaikalia, and the Amur region), the above forms do not exist at all. There, the fauna is represented chiefly by brachiopods with predominant boreal genera (*Tomilia*, *Balakhonia*, *Orulgania*, and others), which essentially were not found in either China or the Urals, nor in any other part of the Tropical region.

By the end of the Early Carboniferous, the Boreal region expanded slightly and embraced almost the entire Siberian craton (Fig. 1). Mixed fauna, including both tropical and boreal elements, was spread in Kazakhstan. Fauna in the Verkhoyansk region was basically composed of boreal elements, though individual specimens of *Gigantoproductus* can still be found there, testifying to the proximity of the boundary.

In the Middle Carboniferous, fauna of the Boreal region was characterized by the lack of fusulinids. The outlines of the region, as presented in Figs. 1 and 4, have been drawn along the line coinciding with the distribution limit of fusulinids, colonial corals, and others. Brachiopods were represented mainly by such endemic genera as *Praehorridonia*, *Jakutella*, *Balakhonia*, *Orulgania*, and *Taimyrella*.

In the Middle Carboniferous, the Boreal region occupied a considerably larger area than in the Early Carboniferous. At the beginning of the Middle Carboniferous, fusulinids were rather widespread in Novaya Zemlya, the New Siberian Islands, Wrangel Island, and in the north of Chukotka, but during the Middle-Late Carboniferous, all tropical elements gradually disappeared there, giving way to boreal ones (Ustritskii, 1970).

The zone of mixed boreal and tropical elements stretched as a rather wide strip along the Arctic, from Wrangel Island to Novaya Zemlya. Mixed faunas and floras are known from Kazakhstan, where a special Kazakhstan floral province has been identified (Vakhrameev *et al.*, 1970). To the east, the boundary between the typically boreal faunas of Mongolia and the tropical faunas of northern China was extremely sharp; mixed faunas and floras are not known here. Equally sharp was the boundary in the Pacific sector, though in the Sette-Daban, rare Middle Carboniferous fusulinids were found, together with typically boreal brachiopods (Abramov and Grigor'eva, 1983). As can be seen in Fig. 1, on the whole, the boundary in this region, unlike the more northern ones, essentially did not change throughout the entire Carboniferous and Permian.

In the Early Permian, the contrast between the biotas of the Boreal and Tropical regions was even more pronounced than in the Carboniferous. Fauna in the Boreal region was surprisingly monotonous. The endemic brachiopod *Jakutoproductus* genus predominated there, and its representatives frequently comprised more than half of the whole biota. This genus associated with the more rare *Attenuatella*, small chonetids and cancrinellids, as well as with the horridonids, confined to coastal facies. At the end of the Artinskian, the Baigendzha ammonoid assemblage became widespread, including the *Paragastrioceras*, *Uraloceras*, and other genera, which were of bipolar distribution or were endemics of the Boreal region. However, the characteristic feature of this region was an absence of several major groups distributed only within the Tropical region.

During the Permian, the boundary of the Boreal region in East Asia remained as sharp and unchangeable as in the Carboniferous. However, its position in the western sector of the Arctic changed substantially. In the beginning of the Permian (Fig. 1), in the Asselian and Sakmarian, fusulinids were widespread in northern Alaska, the Canadian Archipelago, northern Greenland, Svalbard, and the Timan-Pechora region, but by the end of the Artinskian, they disappeared from all of the above areas, which became parts of the Boreal region. However, their fauna occasionally included some groups that are more typical of the Tropical region (for instance, tetracorals).

During the Late Permian, the biotic differences between the Boreal and Tropical regions became utterly profound (Ustritskii, 1970). The Boreal region had no fusulinids, conodonts, colonial corals, echinoids, or a number of other large taxa. Brachiopods, such as *Lyttonia*, *Enteletes*, *Richthofenia*, and others, widespread in the Tropical region, were also absent.

Very typical of the Boreal region were *Inoceramus*-like pelecypods (*Kolymia*, *Atomodesta*, etc.) and finely striated Linoproductidae (*Terrakea*, *Spitsbergenia*, *Cancrinelloides*). Ammonoids became scarce; in the Ufimian, they were represented mainly by the endemic *Tumaroceras* and *Spirolegoceras* genera. By the end of the Kazanian, they disappeared from the majority of the region, remaining only in its margin, in Greenland. The fact that some areas of the Boreal region had negative winter temperatures can be confirmed not only by the character of the deciduous Angara flora, but also by the wide-spread ice-rafted marine sediments in the Verkhoyansk area and in the Northeast of Russia (Ustritskii, 1975).

Because the phyto- and zoogeographic regions in the Carboniferous and the Permian coincided, it is usually possible to reliably define the boundary of the Boreal region. However, there were some considerable deviations in the Late Permian, when the gulfs of the West European Zechstein and the Kazanian Sea of the Russian craton cut deep southward into the limits of the Euramerican floral region. The gulfs were isolated from the Tethys but were closely linked with the Boreal Ocean and were, naturally, inhabited by boreal fauna. In this case, the boundary of the Boreal region shown in Fig. 1 has been drawn according to the data on the flora.

As can be seen in Fig. 1, the Late Permian Boreal region was shaped rather like a regular oval, with the present-day North Pole, which previously lay beyond

its limits, situated not far from its center. In other words, the position of the Late Permian Boreal region resembled that of the present day.

CORRELATION BETWEEN PALEOBIOGEOGRAPHIC AND PALEOMAGNETIC DATA AND GEODYNAMICS

The biogeographic zonation for the Early Silurian looks rather unusual in terms of the present-day location of the continents. A.J. Boucot and J.G. Johnson ("Atlas of Paleobiogeography," 1973) have identified only two main biochores for this time, naming them the Mal'vina-Kaffir and Cosmopolitan biogeographic regions.

The former of the two embraces the larger part of the Southern Hemisphere, including Africa and South America (there is no reliable data on the Antarctic Continent). The absence of graptolites, reef-building, and other groups typical of the tropics, leaves no doubt of the fact that the Mal'vina-Kaffir region corresponded to the southern circumpolar climatic region of the Earth, or the Notal biogeographic region. This view is shared by all scientists, who investigated the Silurian biogeography on the scale of the entire Earth ("Paleobiogeograficheskii Atlas...," 1979; "Paleontologiya...," 1981; "Atlas of Paleobiogeography," 1973). The fauna of the Cosmopolitan region, which included almost all of the Northern Hemisphere and Australia, was typically tropical at the onset of the Silurian. As was mentioned above, it was only since the late Llandoverian that the *Tuvaella* fauna, which was the progenitor of the Boreal region, began to segregate within its limits, from Tuva in the west to the Amur region in the east.

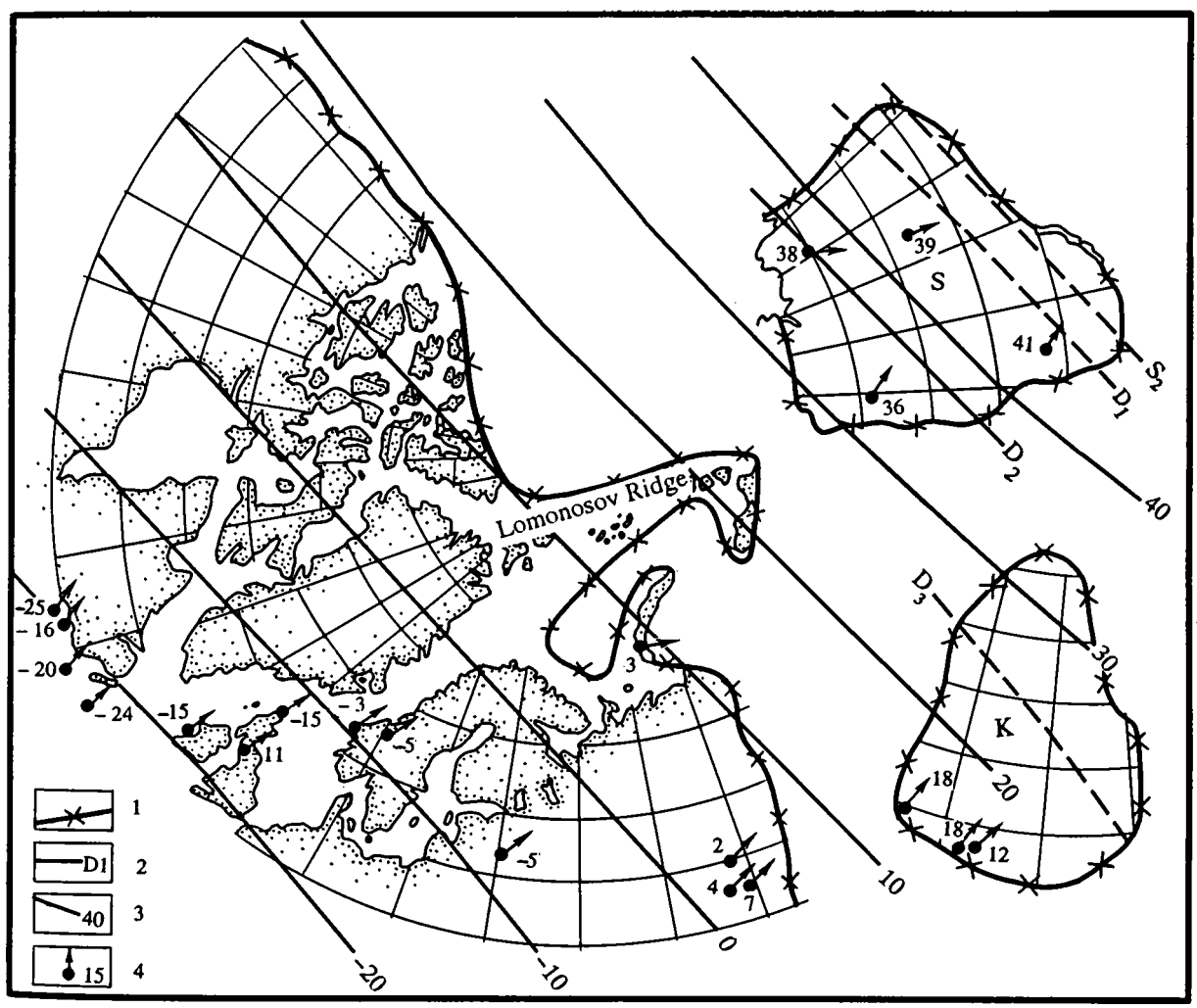

Fig. 2. *Boundaries of the Boreal biogeographic region in the Late Silurian and Devonian, superimposed on A.N. Khramov's Middle Devonian reconstruction based on paleomagnetic data. S – Siberian craton, K – Kazakhstan. 1 – boundaries of continents, 2 – boundaries of the Boreal region, the letter symbol shows the time for which the boundary is drawn; 3 – paleolatitudes according to paleomagnetic data; 4 – paleomagnetic declinations and paleolatitudes for certain points (Middle Devonian).*

Thus, the Silurian deposits throughout almost the entire Southern Hemisphere contained subantarctic fauna, while they contained tropical fauna in almost the whole Northern Hemisphere, the boreal fauna being practically absent. This situation looks paradoxical and can hardly be explained logically from the fixism positions without admitting essential horizontal drift of the continents. At the same time, this situation agrees well with the conceptions of mobilism, based on paleomagnetic, paleoclimatic, and other data.

It is common knowledge that according to this data, South America, Africa, Arabia, and Hindustan occurred further south in the Early and Middle Paleozoic than they do today. Together with Antarctica and Australia, they used to constitute an integrated supercontinent of Gondwana ("Paleomagnitologiya," 1982; "Paleontologiya...," 1981; Khain and Bozhko, 1989). The coldest circumpolar part of that supercontinent embraced South America and West Africa. In the reconstruction of Gondwana, the outlines of the Mal'vina-Kaffir region form an almost regular circle, in which southwest Africa proves to be the center. The northern boundary of this region coincided approximately with 40° south paleolatitude (Silurian!). Australia, though it was part of Gondwana, was the most removed from the pole ("Paleomagnitologiya," 1982). Its northern part almost reached the equator, and therefore, the presence of tropical fauna was quite natural there.

The continents, which today occur in the temperate and high latitudes of the Northern Hemisphere, were situated in the tropics in the Early Paleozoic. The whole of the polar and circumpolar part of the Earth's Northern Hemisphere north of 30°N was occupied by the ocean. The bulk of its lithosphere, together with sediments and fauna, was absorbed in the Benioff zones, which evolved in the frontal part of the continental drift northward.

In the Silurian, the Siberian craton was turned in such a way that the Baikal region was its northern margin, and the Enisei region was its southern margin. In the Devonian, Taimyr (or, to be more exact, its northern part) was separated from the Siberian craton by the ocean (Fig. 2), which closed in the Late Paleozoic simultaneously with the Uralian Ocean. The Taimyr-Severnaya Zemlya block at that time was either a separate microcontinent or was adjacent to the Lomonosov Ridge and Franz Josef Land. Judging from paleomagnetic data ("Paleomagnitologiya," 1982; Ustritskii and Khramov, 1989), in the Ordovician, the continents, now lying in the Northern Hemisphere, began their general northward drift. At that time, the northernmost position was assumed by the areas of Tuva, Transbaikalia, and the Amur region. That is why it was here that relatively severe climatic conditions arose during the Silurian. As a result, the most thermophilic groups gradually became extinct, and fauna became generally poor. This was the most characteristic feature of the Boreal region.

In the Devonian, the continents continued their gradual northward drift. Fig. 2 shows a reconstruction

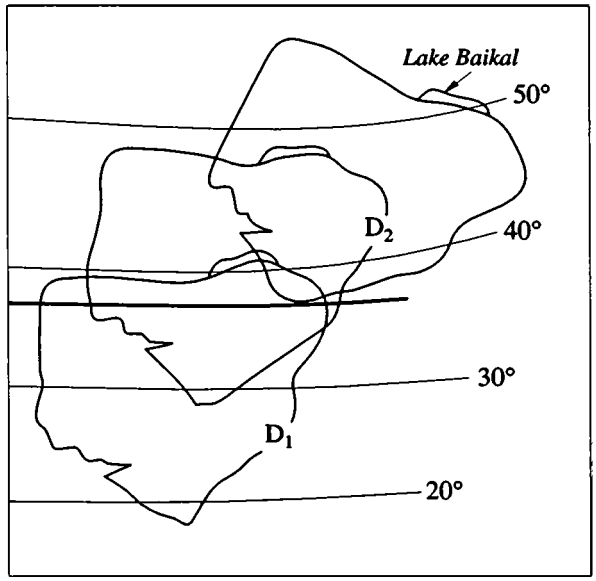

Fig. 3. *Position of the Siberian craton in different epochs of the Devonian. The heavy line shows the Boreal-Tropical boundary. In the Early Devonian, almost all of the craton occurred within the Tropical region; by the end of the Devonian it completely passed into the Boreal region.*

made by A.N. Khramov of the Middle Devonian according to paleomagnetic data and, superimposed on it, the boundary between the Boreal and Tropical biogeographic regions. The boundary in the Late Silurian and Early and Late Devonian is shown by dashed lines. The figure clearly shows that the Devonian boundary between the regions shifted regularly.

During the Late Silurian and Early Devonian, the Boreal region embraced only the northern (in present-day coordinates, the southeastern!) part of the Siberian craton. However, by the end of the Devonian, the Boreal region already included not only the entire Siberian craton, but also a part of Kazakhstan, i.e., the boundary of the region shifted by ca. 25°. It still remained parallel to the Devonian paleolatitudes and crossed the parallels of today at an angle of 45° - 60°. The latter circumstance does not permit attributing the changed position of the boundary to only the climatic warming of the time. Such a regular shift of the boundary logically resulted from the northward drift of the continents. Figure 3 demonstrates the position of the Siberian craton in the Early, Middle and Late Devonian. As one can easily see in the figure, the position of the boundary, and hence the dimensions of the Boreal region, essentially did not change. Throughout the Devonian, the boundary coincides with 35° - 40°N. (To be more precise, the deviations caused by the climatic changes cannot to be identified either by the geomagnetic or by the paleobiogeographic methods.) At the same time, the Siberian craton itself, in terms of its position with respect both to the boundary of the Boreal region and to the parallels, shifted northward by 20° - 25° and during the Devonian moved from the Tropical to the Boreal region.

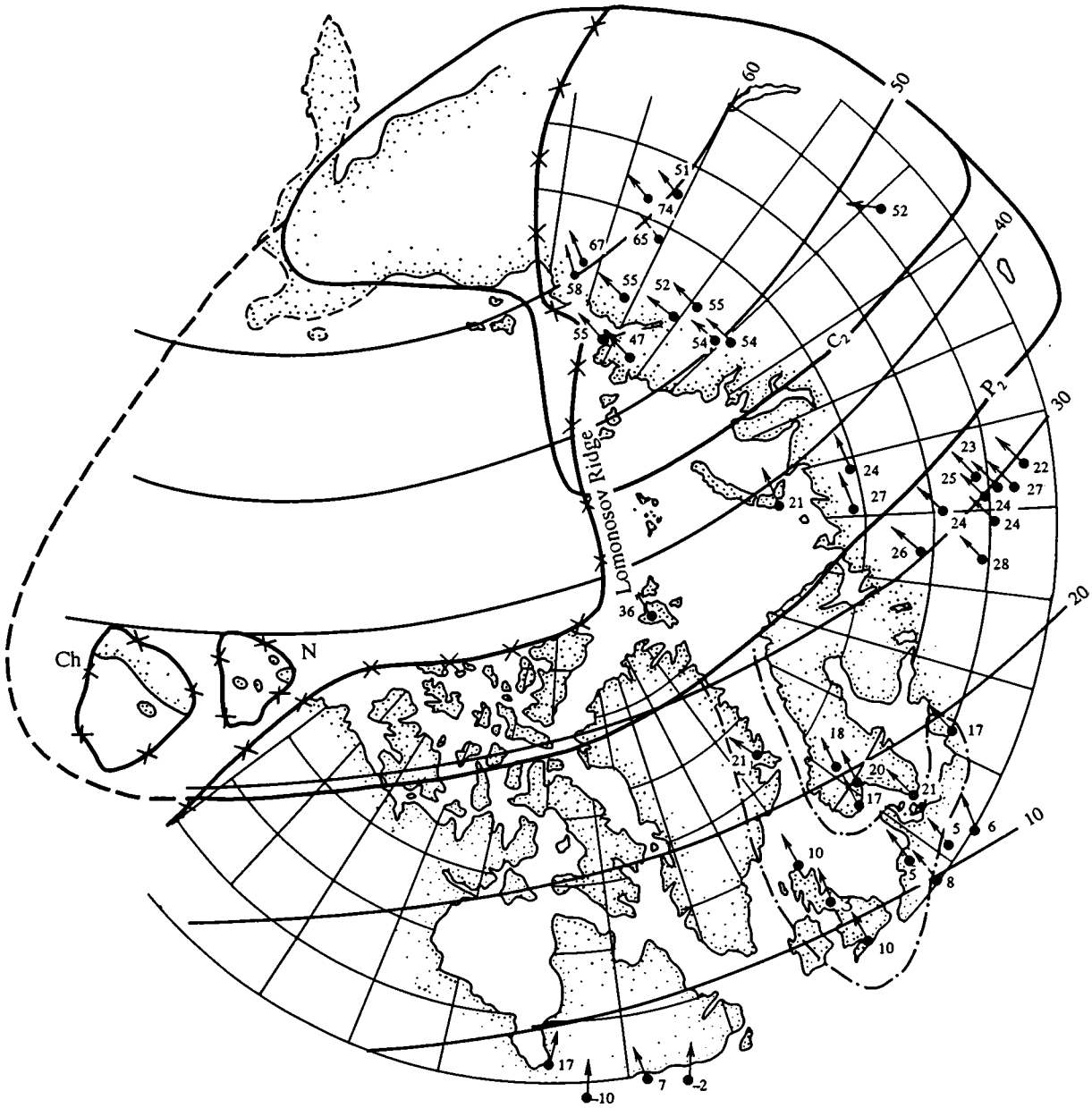

Fig. 4. *Boundaries of the Boreal biogeographic region in the Late Permian and Middle Carboniferous superimposed on A. Khramov's Late Permian reconstruction based on paleomagnetic data.*
Dot-and-dash-line shows the boundaries of the Late Permian Zechstein Sea. Ch – Chukotka-Wrangel microcontinent, N – New Siberian microcontinent. For explanation see Fig. 2. Points of declinations and paleolatitudes are shown for the Late Permian.

During the Carboniferous and Permian, the general northward drift of the continents continued, and more and more, new parts of continents and individual microcontinents drifted from the Tropical to the Boreal climatic region. The northward drift was accompanied by the turn of the Siberian craton. A comparison of Figs. 2 and 4 shows that from the Silurian to the end of the Late Paleozoic, it turned clock-wise by almost 40°. In the Middle-Late Carboniferous, Euramerica (the Atlantic Ocean did not exist then), Kazakhstan, and the Siberian craton were united into a single continent, Laurasia. Therefore, the position of the Boreal region in the Late Paleozoic is considered in a new reconstruction (Fig. 4).

Figure 4 shows the position of the continents and the boundary of the Boreal region during the Late Permian. In addition, a dashed line represents the position of the boundary for the Middle Carboniferous. The boundary between the regions in the Middle Carboniferous and

Permian, like in the Devonian, proves in the huge continent of Laurasian to be parallel to the latitudes established by paleomagnetic data. It passes at approximately 35 - 40° N. (It should be remembered that Fig. 4 shows the position of the continents not for the Middle Carboniferous, but for the Late Permian.) The Boreal region boundaries in the Middle Carboniferous and Late Permian did not coincide because of the further general northward drift of the continents, rather than climatic warming. In this respect, the situation was identical to that characterized for the Devonian (Figs. 2, 3).

After Laurasia arose, it continued the general northward movement. In the Middle Carboniferous, the first to reach 60°N was the Omolon massif, where ice-rafted marine sediments began to sporadically appear (Ustritskii and Yavshits, 1971). In the Late Permian, a part of the Siberian craton, adjacent to the Verkhoyansk region, also reached 60°N. During the Gizhiga cooling, which occurred, most likely, in the Kazanian, the continuous distribution range of ice-rafted marine sediments involved the entire Verkhoyansk region and the Omolon massif (Ustritskii, 1975). In all other time spans, the climate, most likely, used to be warmer, and any traces of sea freezing and ice-rafting of clastic material cannot be found.

Laurasia occupied a considerable part of the Northern Hemisphere since the Late Paleozoic. However, at that time, almost all of the circumpolar area north of 65°N was still occupied by the ocean (Zonenshain and Natapov, 1987; Ustritskii and Khramov, 1987, 1989). Within the limits of the ocean were the geoblocks of the New Siberian Islands, northern Chukotka, Wrangel Island (perhaps, together with the Chukotka dome), and northern Alaska. In the Early Carboniferous, they were still situated within the limits of the Tropical climatic region, i.e., south of 40°N. During the Late Paleozoic, they were drifting northward, as testified by the fact that tropical fauna was successively replaced by boreal fauna. In the middle of the Bashkirian, the very last fusulinids disappeared from the New Siberian Islands; in the middle of the Moskovian, they vanished from Northern Chukotka, and in the Artinskian, from Northern Alaska. Fusulinids everywhere gave way to typically boreal fauna.

The fact that fusulinids in all geoblocks under consideration disappeared at different times and that, therefore, their movement from the Tropical to the Boreal region was not simultaneous, does not quite agree with the ideas advanced by L.P. Zonenshain and L.M. Natapov (1987), who assert that during the Late Paleozoic, all of the above geoblocks were parts of an integrated continent, which these scientists named Arctide. It seems likely that during the Late Paleozoic, the New Siberian Islands, Wrangel Island, and northern Alaska were independent microcontinents, which joined Laurasia in the Mesozoic, when the Anyui Ocean, which in the Late Paleozoic directly continued from the Pacific Ocean (or, to be more exact, was part of it), became closed (Fig. 4).

CONCLUSIONS

1. Considering paleobiogeography, one finds that the main factor determining the regularities in the distribution of the Earth's fauna during the Paleozoic, and during all other subsequent epochs, was the climatic zonality. It means that the principle of identifying the basic biogeographic units corresponding to climatic zones can be applied to the entire Paleozoic. At the present-day stage of our knowledge about the Paleozoic, it seems most feasible to restrict our identification to that of the Boreal, Tropical, and Notal regions (superregions). Smaller biochores (provinces) can be specified within their limits. The life-span of such provinces may be governed by the actual paleogeographic features of one or another period. The most fundamental difference between smaller biochores and regions is the fact that the regions, except the Boreal region, which arose during the Silurian, existed throughout the Phanerozoic, irrespective of the paleogeographic features of the particular period (Ustritskii, 1986).

When examined further, the fact that at the boundaries of the regions, there are wide zones of mixed tropical and boreal, or tropical and subantarctic, elements, may permit us to classify these zones as independent biochores (belts) and to start identification of biogeographic belts, i.e., to create an integrated system of biogeography for the entire Phanerozoic.

2. The regions are most clearly distinguished by flora and by benthic groups of fauna that inhabited mainly the epicontinental shallow seas. I know of almost no benthic faunal finds in abyssal sediments of the circumpolar regions. The distribution range of most planktonic (graptolites) and nektonic (conodonts) groups is confined to the Tropical region. Taking into account these comments, the following features of the Paleozoic Boreal and Notal biogeographic regions may be considered to be their most characteristic ones: (a) the lack of major taxa, which were distributed only within the Tropical region (tabulates, colonial tetracorals, almost all reef-building and planktonic organisms, and others), and the resulting monotony of fauna; (b) the low rank of endemics (usually species and genera; families are found very rarely); (c) fauna including large numbers of taxa with bipolar distribution.

3. The dimensions of the Boreal biogeographic region during the Middle and Late Paleozoic remained much smaller than those of the Notal region (Gondwana). During the Late Paleozoic, the latter embraced almost all of the Southern Hemisphere of the Earth, and even Hindustan, which was already situated in the Northern Hemisphere. Throughout the Paleozoic, the dimensions of the Notal region practically remained constant, though the position of its boundaries on continents did change as a result of Gondwana's general displacement with respect to the pole. The huge dimensions of the Paleozoic Notal region on the present-day Earth resulted from the disintegration of Gondwana and from the northward drift of its continents, but

it had already taken place in the Mesozoic ("Paleontologiya...," 1981; Khain and Bozhko, 1989).

Unlike the Notal region, the Boreal biogeographic region was steadily expanding in size throughout the Middle-Late Paleozoic. This was because of the general growth of the area under the epicontinental seas with their boreal fauna on the continents, which were moving from the Tropical to the Boreal climatic region.

4. A joint analysis of paleobiogeographic and paleomagnetic data shows that they are in nearly ideal agreement. The most essential biotic changes, determining the boundaries between the regions in all periods of the Paleozoic, occurred between 30° and 40°S and N. Because it is at these latitudes that the Carboniferous and Permian replacement of evergreen plants by deciduous ones can be noted, it seems likely that these boundaries corresponded to the borders of the zones with stable negative winter temperatures. It is very likely that this conclusion can also be applied to all younger periods of the Paleozoic.

I express my sincere gratitude to A. N. Khramov for having supplied the latest paleomagnetic data and for his constant advice and consultations, without which my research could never have been accomplished.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 80 - 86.
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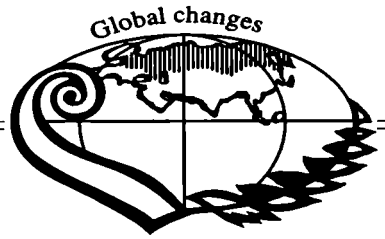

Evolution of the biosphere

The Amka Cooling and Floral Evolution at the Early-Late Cretaceous Boundary

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Manuscript received March 20, 1992

Abstract – At the end of the Albian, the Amka extreme cooling has altered the floral communities from mesophytes to cenophytes. The thermophilic taxa did not survive the cooling peak, and species composition of the relic gymnosperm genera changed drastically. The Late Cretaceous (Cenomanian) warming triggered the first universal burst in the diversity of angiosperm species.

One of the most spectacular changes in the plant life of the Earth occurred at the Early-Late Cretaceous boundary, when many mesophytic plants became extinct or changed their species composition, and a new group of plants, the angiosperms, acquired wide distribution.

The change in plant life was promoted by cooling at that time, as testified by the fossil mountain floras of the Okhotsk-Chukotka volcanogenic belt, because the changes in climate had the greatest effect on these floras. This cooling was called the Amka extreme cooling, after the flora type locality (the Amka River). The sedimentary deposits have no analogs of the Amka coldest floras from the volcanogenic belt.

The Albian-Cenomanian interval shows several climatic changes: an optimum in the Early Albian, a gradual cooling with an extreme phase in the latest Albian, and a relative warming in the Cenomanian. Moreover, the Early Albian and Cenomanian warmings occurred on a global scale, as confirmed by floral data from the North Pacific and by the changes in foraminiferal assemblages in the South Atlantic, i.e., on the opposite sides of the globe (Krasheninnikov *et al.*, 1990).

The paleotemperature data corroborates these climatic changes. For example, in the basins of the Crimea and the Caucasus, the temperatures (determined mainly from the calcium-magnesium ratio) decrease from 21°C in the Early Albian to 15°C in the Middle Albian and to 11°C in the Late Albian. In the Cenomanian, the temperature rose again (Yasamanov, 1978).

When did the Amka cooling actually take place?

Lying above the Amka flora of the cooling period, the Grebenka type floras in Northeastern Asia correspond to another warming, marked by abundant cycadophytes and the presence of thermophilic *Cycadites*. In the Ginter Formation, their age, according to position relative to the marine strata, corresponds approximately to either the middle or a part of the Early Cenomanian (Pergament, 1978; Lebedev, 1987). The Middle Cenomanian floras of the Crimea, dated by fauna, correspond to the warming (Krasilov, 1984), as does the Late Cenomanian Danvegan flora in western Canada (Bell, 1963; Stott, 1968).

It is suggested that the Grebenka type flora and, consequently, the corresponding warming, could also have taken place in the Late Albian. We must assume, however, that the Cenomanian warming could have started at the very end of the Late Albian; moreover, the presence of thermophilic foraminifers in the South Atlantic indicates that this warming continued until the end of the Cenomanian, but its onset time is obscure because of a hiatus in deposits (Krasheninnikov *et al.*, 1990). How long, then, could a time period of the Late Albian be, when the warming took place? According to Vakhrameev (1988, p.180), "...the larger part of the data indicates that the Late Albian is almost universally marked by lower temperatures."

The heavy prevalence of cold-resistant conifers in the Amka beds and the disappearance of thermophilic cycadophytes signify considerable intensity of the Amka cooling, as it changed the floral community on the vast territories of the Northern Hemisphere, from western Kazakhstan and Yakutia to Northern Alaska. From the Early Albian until the end of the epoch, the share of conifers grew and that of cycadophytes decreased in all areas (Lebedev, 1987). Therefore, the Amka cooling, which occurred between the Early Albian and Cenomanian warmings, like the warmings themselves, should be of a global scale and, therefore, would have left evidence over the larger part of the

Earth's surface. Some of such evidence was cited above.

To define more precisely the age of the Amka cooling, it is important to compare the floras of the Albian-Cenomanian in the Okhotsk-Chukotka belt with those of Northern Alaska, where the age of floras is dated by ammonites, inocerams, and foraminifers from marine deposits, interbedded by flora-bearing beds or substituted by them along the strike (Smiley, 1972). The result shows a coincidence in the development of the main climate-indicating plant groups, primarily conifers and cycadophytes. In Northeastern Asia, the Amka floras have a maximum of conifers (45%), and in Northern Alaska, the middle part of the third floral "zone" (after Smiley), i.e., the end of the Late Albian, contains 50% of conifers. Accordingly, the Amka peak of conifers can also be dated the latest Albian (Fig. 1).

The final answer can be reached through more precise dating of the "conifer peak" using faunal data together with detailed paleobotanical and paleotemperature studies. Though the latter produce rather contradictory results, still a fair amount of data from the Southern Ukraine, Crimea, and Central Asia, including Mugodzhary, indicates that the cooling occurred at the end of the Albian (Yasamanov, 1978) (Fig. 2).

An analysis of the Amka cooling and the amplitudes of the temperature decline shows three stages in the evolution of the Amka flora when conifers were definitely dominant. The lower stage still contains rare cycadophytes, but in the middle stage they disappear. In the upper stage, cycadophytes appear again, giving rise to occasional accumulations. Thus, the maximum cooling falls into the Middle Amka stage; in the early and particularly in the late Amka stages, the temperatures were apparently somewhat higher.

Persistent cooling in Albian time shifts to the south the boundary between the temperate and subtropical floras. The range of the shift indicates that the average temperature on the Earth's surface in the Middle Amka time was reduced by at least 5°C (Lebedev, 1982). If the growth of the latitudinal thermal gradient during climatic coolings is taken into account, then the temperature decrease might have been in the range of 5° to 8°C.

The Amka flora is strikingly different from the underlying and overlying floras by an unusually strong dominance of conifers, implying the onset and the ceasing of the extreme cooling to be sudden. Recently, one of the index forms of the Amka stage, the conifer *Metasequoia*, was found in the underlying layers; this conifer is generally characteristic of cooler climate periods.

Consequently, on the strength of this evidence, the onset of the Amka cooling was fairly rapid but not as sudden as was previously thought, and its change to warming in the Grebenka stage was much more gradual.

The Amka cooling, therefore, is the final stage of a persistent cooling process during the Albian time. But even against this background, it stands out as the peak of temperature decline. It was caused, most likely, by

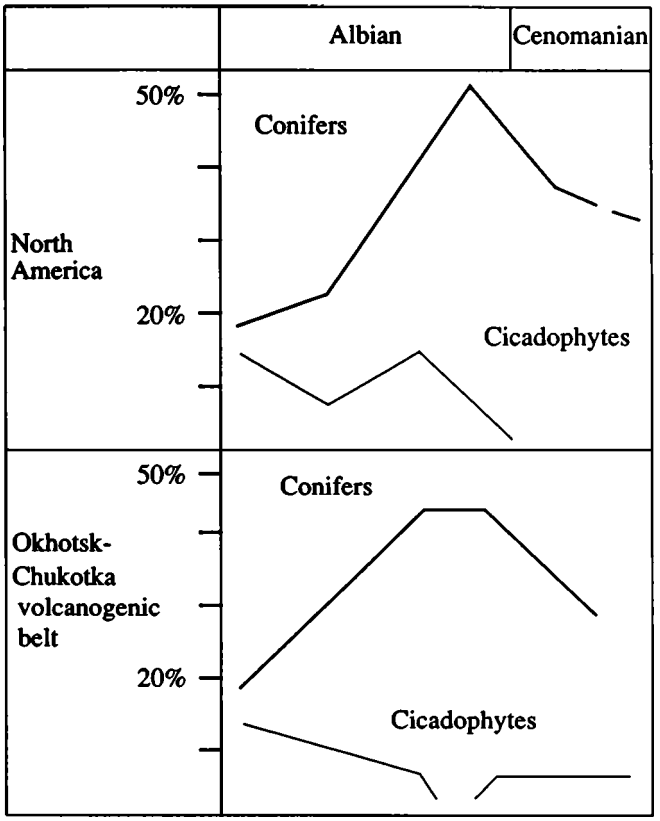

Fig. 1. *Evolution of conifers and cycadophytes of the Middle Cretaceous floras in Northern Alaska and the Okhotsk-Chukotka volcanogenic belt (schematic representation after Smiley (1972) and Lebedev (1987)).*

catastrophic explosive eruptions that contaminated the atmosphere and screened the Earth's surface from solar radiation (Lebedev, 1982; 1987).

The Amka extreme cooling upset the ecology-competition balance and caused the extinction of, or change in, the mesophytic plant groups composition. It also launched the adaptive radiation of cenophytic angiosperms, the group most completely reflecting all climatic changes. Moreover, their active evolution at that time formed general features of the Late Cretaceous vegetation.

To better understand the changes in vegetation, the process should be analyzed in different natural conditions of the Northern Hemisphere, from the zone of temperate vegetation in the north to subtropics in the south, where both the Albian and Cenomanian floras occur. The paleofloral zonation is drawn from V.A. Vakhrameev's data (1988).

The area of Northeastern Asia is considered here as the key region. In the Okhotsk-Chukotka volcanogenic belt of this region, the cooling most drastically affected the continuous succession of the Albian-Cenomanian mountain floras. Moreover, there are rich floras of the Early Albian-Cenomanian in this region's plain habitats. It is important to note that these floras were evolving in a humid climate. All these peculiarities allow us

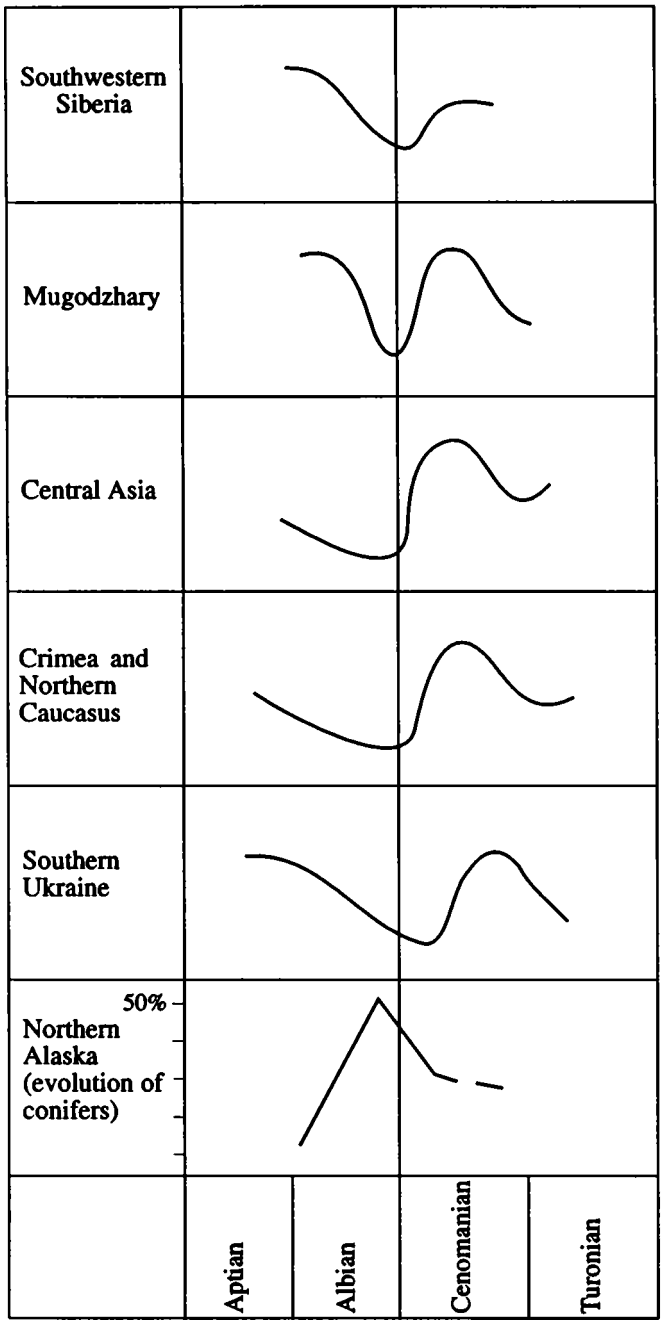

Fig. 2. *Climatic changes in certain regions of the Northern Hemisphere and evolution of conifers in Northern Alaska at the Early-Late Cretaceous boundary (schematic representation after Smiley (1972) and Yasamanov (1978)).*

to make a sufficiently detailed analysis of vegetation readjustment and to establish relations between this process and climate changes.

In the same region, within the North Pacific refugium, several mesophytic gymnosperms have survived the Early-Late Cretaceous boundary; on the rest of the planet they became extinct earlier (Vakhrameev, 1981). Primarily, these are *Nilssonia*, *Phoenicopsis*, and *Sphenobaiera*. The way to survive was different for relics of the plain and mountain habitats, i.e., in a different landscape-climatic environment. For example, *Phoenicopsis* and *Sphenobaiera* disappeared in the plain habitats at the beginning of the Late Cretaceous, whereas in the Okhotsk-Chukotka belt, in the mountain environment, they remained abundant actually until the end of the Cretaceous (Lebedev, 1987). The ferns *Birisia* and *Arctopteris*, typical of Northeastern Asia, appeared in the second half of the Early Cretaceous and also survived that critical point.

The first angiosperms appeared in Northeastern Asia in the early Albian. Most of the species (22) were discovered in the middle coarse-grain part of the Buorkemus Formation (Samylina, 1960), among them, *Cercidiphyllum potomacense*, *Crataegites*, *Dalbergites*, *Celastrophyllum*, *Dicotylophyllum* spp., and others. They are represented by parvifoliate forms with undeveloped venation. The leaves of the Middle Albian angiosperms belong to this type.

The cenophytic angiosperms, however, appeared only in the Late Albian. The first was *Menispermites*; somewhat later, *Trochodendroides*, *Platanus*, and *Dalembia* became the common components of the Late Cretaceous floras of the region (Fig. 3).

In the Cenomanian, the Grebenka floras show a burst of angiosperm species diversity in response to the warming. During this period, *Menispermites*, *Platanus*, *Dalembia* become increasingly numerous. For the first time, *Magnolia* and *Viburnum* appeared and, as yet known only at the Grebenka level, *Grebenkia* and *Sorbites*. A number of other plants were also present: *Leguminosites*, *Dalbergites*, *Myrtophyllum*, *Zizyphus*, *Araliaephyllum*, and others.

The ferns *Birisia*, still numerous at the beginning of Late Albian time, are not found in the Amka beds; they appear again during the Cenomanian warming, but in a different assemblage. If in the early Albian, the ferns were represented by *B. acutata*, *B. alata*, *B. onychioides*, and others, in the Cenomanian, they included *B. oerstedtii* and *B. elisejevii*. The same is true of *Arctopteris*, which is also found in the cooler Amka floras. The representatives of the genus *Coniopteris*, so diverse in the early Albian (9 forms), are rare in the Cenomanian Grebenka floras, which have many thermophilic *Gleichenia*, *Birisia*, and *Onychiopsis*.

Most of the cycadophytes typical of the Early Cretaceous did not survive the transition to the Cenomanian. In the middle of the Albian, the genera *Neozamites* and *Anomozamites* disappeared, and the relic genera changed their composition. The composition of conifers altered at the beginning of the Albian; in the middle of Late Albian, the genus *Metasequoia* appeared and still exists.

Thus, in the Late Albian, three angiosperm cenophytic genera *Menispermites*, *Platanus*, and *Trochodendroides*, appeared and developed further in the Late Cretaceous. In the Cenomanian, they are joined by *Magnolia* and *Viburnum*. The formal genera *Dalembia*, *Grebenkia*, *Sorbites*, and others, which appeared at that

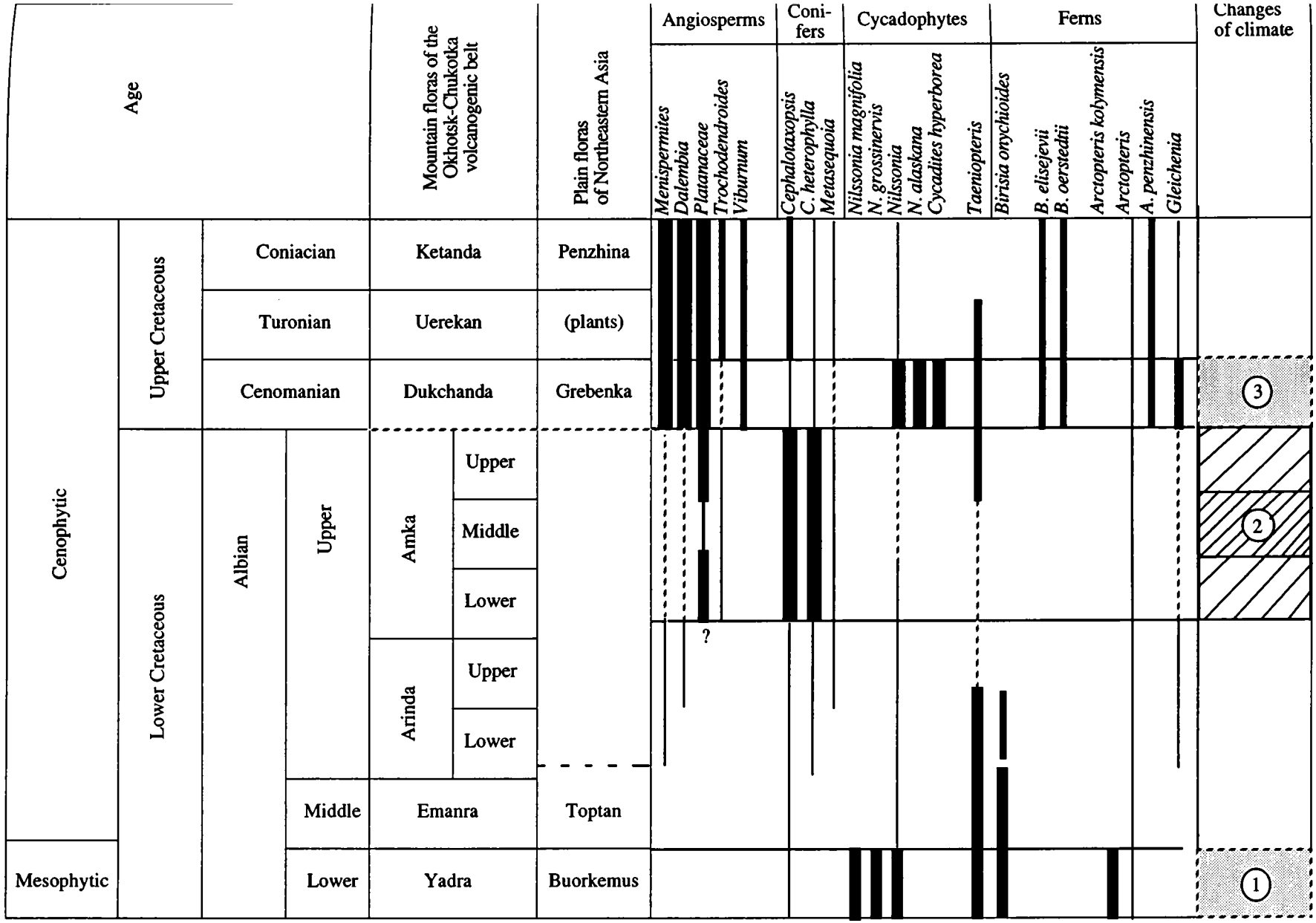

Fig. 3. *Distribution of individual taxa of Cretaceous plants against the background of climatic changes in high latitudes of Northeastern Asia at the Early-Late Cretaceous boundary (compiled from the data of Samylina (1974), Lebedev (1987), and new discoveries). The thickness of lines corresponds to the frequency of occurrence or taxon diversity; striation shows cooling, and dots show warming; from below upward: (1) the temperature optimum in the Early Albian, (2) the Amka extreme cooling at the end of the Albian, and (3) relative warming in the Cenomanian.*

time, are quite peculiar and will probably be classified later as separate families.

In contrast to angiosperms, the cycadophytes have suffered the elimination of genera, whereas the relic genera, e.g., *Nilssonia*, changed their species composition. This is, most likely, also true of the Late Cretaceous *Phoenicopsis* and *Sphenobaiera*, which in volcanogenic formations, unfortunately, lose their cuticles, thus making differentiation of species difficult.

The Amka cooling also caused a sudden change in the mesophytic groups of plants at the Early-Late Cretaceous boundary. The most striking example is provided by thermophilic cycadophytes and ferns that did not survive the cooling; after the cooling, they are represented by new species (Fig. 3). It is worth mentioning that most species of ferns (as also of cycadophytes), which existed during the early Albian temperature optimum (about 60 species) (Samylina, 1974), did not survive the cooling at the transition to the Cenomanian.

The Amka cooling, in other words, caused a sudden change in the species and genera composition of mesophytic groups of plants and destroyed the existing mesophytic plant associations. The cooling, however, essentially did not reduce the adaptive radiation of cenophytic angiosperms and strengthened their hold in vegetation cover. Consequently, the basic changes in vegetation took place in the Late Albian, apparently, in its second half.

Similar tendencies are observed in the evolution of angiosperms and other plant groups in the Albian-Cenomanian floras of western Kazakhstan, where the Middle Albian, the Late Albian, and the Cenomanian floras are recognized. The latter two vegetated in the ecotone zone along the southern margin of the Siberian-Canadian paleofloral province (Vakhrameev, 1952; 1988).

As in the north, the cenophytic angiosperms *Menispermites* and *Platanus* appeared here in the Late Albian. However, in this region, *Platanus* representatives are more diverse and abundant, and their imprints reach 90% of the total number of imprints (Vakhrameev, 1952). In the Cenomanian, they are supplemented by *Magnolia* and others.

Among the 17 genera of the Late Albian are noted *Menispermites*, *Platanus*, *Leguminosites*, *Dalbergites*, *Zizyphus*, *Sterculia*, *Araliaephyllum*, and others. The Cenomanian deposits contain *Magnolia*, *Menispermites*, *Platanus*, *Anacardites*, *Paliurus*, *Zizyphus*, *Sterculia*, and other angiosperms; they make up 70% of the total number of species.

The Middle Albian angiosperms are small-leaved, with undeveloped venation, such as *Nelumbites*, *Leguminosites*, *"Cercidiphyllum" potomacense*, *Celastrophyllum*, and others.

It is significant that most of the mesophytic ferns and all of the cycadophytes did not survive the transition to Cenomanian time. In the Middle Albian, the ferns were still diverse (43%); among them were *Adiantopteris*, *Phlebopteris*, *Weichselia*, various *Gleichenia*, and trunks of *Tempskya*. In the Late Albian, the ferns were less diverse; in the Cenomanian, they comprised less than 10% of the total number of species.

In the Middle and Late Albian, the cycadophytes *Nilssonia*, *Otozamites*, and *Zamites* (?) were still present; in the Cenomanian, they were already extinct.

It is interesting to note that the evolution of angiosperms and cycadophytes also has tendencies similar to subtropical floras; for example, the Potomac Group in the United States contains floras of the Aptian, Albian, and Cenomanian, briefly characterized below (Berry, 1911, 1916; Doyle and Hickey, 1976; Fontaine, 1889).

The Aptian angiosperms (the Patuxent Formation, the bottom of the Potomac Group) are ancient forms with undeveloped venation and indistinct rank differentiation of veinlets of different rank. They are referred to the formal genera *Acaceaephyllum*, *Proteaephyllum*, *Rogersia*, *Plataginopsis*, and others.

In the Albian (the Potapsco Formation, the top the Potomac Group, i.e., the Middle, or probably Late Albian), the diversity of the angiosperms increases and their venation is more organized. At that time, *Menispermites* and *Nelumbites*, *Araliaephyllum*, *Alismophyllum*, *"Populus" potomacensis*, and *Sapindopsis* spp. appeared.

In the Cenomanian (the Raritan Formation), the diversity of the angiosperms was even greater. The plants, also known in Siberian floras, *Magnolia*, *Menispermites*, *Platanus*, *Viburnum*, *Leguminosites* and subtropical *Debeya* (*Dewalquea*), *Liriodendron*, *Andromeda*, and others (Berry, 1916) comprised a part of the flora. This period yielded an abundance of imprints of narrow and entire leaves.

Almost as diverse are the angiosperms in (mostly) Cenomanian deposits of the Dakota Formation in the central part of the United States. *Magnolia, Menispermites, Platanus, Protophyllum, Aspidophyllum, Viburnum, Araliaephyllum*, and subtropical *Liriodendron*, *Liriophyllum*, and others are found here.

The ferns are still numerous in the Aptian: *Ruffordia*, *Onychiopsis*, various *Acrostichopteris*, and the trunks of *Tempskya*. In the Albian and particularly in the Cenomanian, however, the diversity of ferns notably decreased. Among the Albian cycadophytes, *Ctenopteris* and *Zamiopsis* were eliminated, and *Nilssonia* was not observed. In the Cenomanian, the relics are represented by *Williamsonia*. The Cenomanian deposits of the Dakota Formation do not contain *Nilssonia*, and *Encephalartites* and *Williamsonia* are relics.

The Aptian conifers include *Nageiopsis*, *Sphenolepis*, *Athrotaxopsis*, and *Sequoia*. The Albian contains *Araucarites* and *Pinus*; the Cenomanian was marked by the appearance of *Protophyllocladus* and *Libocedrus* (*Androvettia*); *Frenelopsis* still existed at that time.

It is important to note that despite many differences, there are still similar tendencies in the evolution of angiosperms (and of other plants) at the Early-Late Cretaceous boundary in all of the zones of the Northern Hemisphere, ranging from the subtropical to the high northern latitudes.

The angiosperms of the Early and Middle Albian (in the south, the Aptian must be included as well) are small-leaved with undeveloped venation; sometimes, they have indistinct differentiation of veinlets of different rank. In the North Pacific refugium, this ancient group of angiosperms probably survived until the end of the Albian and coexisted with the developed angiosperm forms. Such leaves were found in association with *Platanus* leaves in the lower Amka beds.

Reliable finds of cenophytic *Menispermites*, *Platanus*, and *Trochodendroides* in the north are only as old as the Late Albian. *Platanus* is more numerous in the southern regions.

At the beginning of the Late Cretaceous (in the Cenomanian), the onset of the warming caused a universal burst in the diversity of angiosperms, and the already existing *Menispermites* and *Platanus* were joined by *Magnolia* and the rather numerous *Viburnum*. Simultaneously, genera specific to either only northern or southern floras appear at that time.

Other Cenomanian floras, located in different natural zones of the Northern Hemisphere, have a similar composition of angiosperms: the Peruč flora, western Europe; the Late Cenomanian Dunvegan flora, western Canada; and the Melozi-Kaltag flora in Alaska. These floras, together with the large leaves of the *Platanus* assemblage, contain certain genera of small entire and narrow leaves. Many of these floras have relic cycadophytes, which are more numerous in the North Pacific refugium, where they survived almost until the end of the Late Cretaceous.

My opinion is that the succession of stages discussed above is a result of changes in the climate on a global scale, mainly, the extreme cooling at the end of the Albian and the Cenomanian warming.

As the data on Northeastern Asia shows, the Amka cooling caused a drastic change in composition of cycadophytes and ferns (Fig. 3) and, consequently, destroyed the then-existing mesophytic plant associations. As was shown above, similar tendencies in the evolution of cycadophytes and ferns are likewise evident in subtropical floras. Therefore, if the tendencies in the evolution of separate groups of plants are similar in different natural zones of the Northern Hemisphere, then obviously their causes should be of a global scope, such as the Amka extreme cooling in the second half of the Late Albian. The effect of the cooling was naturally less on the southern floras.

The Cenomanian burst in the diversity of species of angiosperms was also associated, most likely, with the warming, also on a global scale, and can be traced in both hemispheres (Krasheninnikov *et al.*, 1990).

By the end of the Albian, the conifers in the north became more numerous toward the cooling peak, whereas in the south, this trend is not so obvious because of the difference in composition of subtropical conifers (Lebedev, 1987).

Therefore, the character of vegetation changed drastically in a relatively short period of time (the second half of the Late Albian). In the south, the plane-tree forests (the ecotone zone) and the forests of various angiosperms (the subtropical zone) had already come into existence by the Late Albian. Since the beginning of the Late Cretaceous, the angiosperms for the first time dominated in floral composition over other plant groups.

According to the evidence given above, the boundary between the mesophytic and the cenophytic must be at the base of the Late Albian, as postulated by V.A. Vakhrameev (1981) and V.A. Samylina (1974).

In other words, the beginning of the cenophytic most likely coincides with the first appearance of angiosperms whose venation approached the modern type; these angiosperms later became the common components of the Late Cretaceous floras.

Sometimes the lower boundary of the cenophytic is attributed to the base of the Aptian, based on the first appearance of angiosperm leaves. In this case, however, it divides the Neocomian and the Aptian – Middle Albian floras, which contain as dominant the same mesophytic plant groups. Moreover, with an accumulation of data, one can expect more ancient angiosperms to be discovered.

Thus, to conclude, the changes in vegetation at the Early-Late Cretaceous boundary were caused by the Amka cooling which lasted approximately 3 - 5 m.y. in the second half of the Late Albian. The cooling was an event of global scale, with an average temperature drop of 5 - 8°C. It particularly affected the composition of the Albian-Cenomanian floras of higher latitudes and primarily, the mountain floras of the Okhotsk-Chukotka volcanogenic belt. The Amka cooling initiated a considerable lateral and vertical migration of vegetation and destroyed mesophytic plant associations by disturbing the ecological balance. The thermophilic taxa did not survive the cooling peak, and the composition of relic genera in refugia changed cardinally. At the very beginning of the Late Cretaceous, in Cenomanian time, the global warming triggered a burst in angiosperm diversity; since that time, these plants have dominated over other plant groups in abundance.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 85 - 92.
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Evolution of the biosphere

Paleocenophytic to Neocenophytic Transition: a Manifestation of Global Change in the Ecosystem at the Eocene-Oligocene Boundary

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Manuscript received April 14, 1992

Abstract – One of the most important changes in Cenozoic ecology at the time of the Eocene to Oligocene transition is analyzed. The cooling of the climate in the middle of the Late Eocene caused a change in climax forest communities of the Late Eocene, being coeval with the regression of epicontinental seas from high latitudes, as is shown by studies in regions of the Northern Hemisphere (northeastern Russia, the Far East, Armenia, Kazakhstan, Western and Central Europe, North America). The alternation of humid and arid climatic phases of climate in the Early Oligocene accelerated the destruction of the old and the establishment of the new zonal types of vegetation. During the new cooling-stage in the Late Oligocene, the warm-temperate mesophilic vegetation of the Turgai ecotype was eventually formed in the middle latitudes of Eurasia and North America.

The geological history of temperate climatic zones of the Northern Hemisphere in the last half of the Paleogene reveals two time intervals with stable ecosystems lasting no less than 5 to 7 m.y. each. They were characterized both by a relative stability of the environment over vast areas of the lowlands and epicontinental seas, and by the rather constant biotic background accompanied by monotonous phytolandscapes. These two intervals can be easily related to the Late Eocene and Late Oligocene epochs. It is by far more difficult to reconstruct the floral composition and the vegetation type in the intermediate stage of stable ecosystems, because the phytolandscapes became mosaic, and the flora itself became mixed because of the gradual absorption of deciduous mesophilic components and the coeval disappearance of evergreens. V.S. Kornilova suggested naming the Kazakhstan floras of this type the "Shintuzsai" floras.

The first interval, often called paratropical, was the time of extensively developing forests. The main forest-forming plants were Lauraceae and Fagaceae in combination with Pinaceae and Taxodiaceae, particularly in the northern regions. Their main ingredients were Araliaceae, Anacardiaceae, Altingiaceae, Platanaceae, Hamamelidaceae, and Nyssaceae. Large areas were covered with riparian and coastal forests containing the ferns *Blechnum* and *Pteridium*, palms, and Myricaceae. The presence of formal genera inherited from the preceding Early Paleogene flora was another characteristic feature of the Late Eocene floras.

An equally extensive development of homogeneous flora and forest vegetation took place during the second time interval; the flora, of the so-called Turgai type according to A.N. Krishtofovich, was temperate-warm, mesophilic, and deciduous. The leading forest-forming group, besides Taxodiaceae and a wide spectrum of Pinaceae, included Amentiflorae (Fagaceae, Inglandaceae, and Betulaceae; riparian formations contained Salicaceae and Myricaceae). The broad-leaved plants were accessory and comprised Ulmaceae, Cercidiphyllaceae, Aceraceae, Tiliaceae, and others, as well as the arboreal and shrub Rosaceae. Formal genera are no longer present in the Late Oligocene floras.

Both vegetation types inherited one another spatially in North America, Western and Central Europe, the Russian Plain, and Kazakhstan. Inspite of the nearly equal water supply, these forests, separated by a time interval of about 5 Ma, had distinctly different thermal regimes: the mean annual temperatures differed by not less than 8 - 10°C.

Such a significant decline of mean annual temperatures in a rather short period of time is as yet known to occur in the Late Phanerozoic only at the boundaries between the Cretaceous and the Paleogene and between the Pliocene and the Pleistocene. The cause-and-effect relations and the consecutive course of events relevant to these boundaries were repeatedly debated. A corresponding reconstruction can be suggested for the time-limit under discussion, as is shown in Table 1.

Table 1. Succession of events in geology and ecosystem at the Eocene-Oligocene boundary in the middle latitudes of the Northern Hemisphere

1. ?	2. A change of the salt-stratification of oceanic water layer to the thermal one. Formation of the Circum-Antarctic current; the thermal isolation of Antarctica		3. The onset of Antarctic glaciation. Large-scale regression
4. Partial exsiccation of epicontinental seas. Drying up of marine meridional estuaries and straits connecting low-middle lati tudes with the Polar Region (the West Siberian Sea, Turgai Strait, the estuary of Mississippi north from the Bay of Mexico)		5. Global cooling	6. The growth of the thermal gradient (Pole-Equator) (an increase in latitudinal climatic differentiation)
7. Cardinal changes in the ecosystems			8. Transition from paleocenophytic to neocenophytic. Deterioration of the old zonal types of vegetation and formation of new types in the middle and high latitudes. The characteristic feature of middle latitudes is frequent alternation of humid and arid phases of climate which affects the destruction and remolding of old coenoses.

The planetary changes were wider than those listed above. Since the Oligocene, the acryogenic climate of the Mesozoic-Early Cenozoic Earth changes to cryogenic; the marine and continental biotas notably change their components, and herbaceous coenoses actively develop in the vegetation cover of the planet.

The transitional period between the two stable ecosystems is discussed here in the light of changes in the Early Oligocene flora and vegetation of the ecotone zone lying between the Boreal and the Tethyan floral regions. In the middle latitudes of Eurasia, this ecotone was not only the boundary between the main climatic zones, the warm-temperate and subtropical, but was also, for example, in Kazakhstan, a transition area from the humid to arid regions. Hence, the Lower Oligocene floras of regions located in the temperate and subtropical areas far away from these ecotone zones should be discussed for comparison (northeastern Russia, the Far East, Armenia). Our conclusions are based on a series of reference sections of the Eocene-Oligocene timespan, which we have studied personally.

Several of the studied flora-bearing sections are located in East Russia (Rarytkin Range, Koryak Highland, northwest Kamchatka, Sakhalin, Sikhote-Alin', South Primor'e). Other sections are in Kazakhstan and the Crimea-Caucasus region (Zaisan and Turgai basins, northern Aral area, Ustyurt, northern Caucasus, Armenia).

We will concentrate our discussion on changes in the Late Eocene-Oligocene flora and climate in different regions of the Northern Hemisphere, with particular attention to horizons of transitional Lower Oligocene floras.

The Far East and Northeastern Russia

In the Paleogene, the middle latitudes of the peri-Pacific region were located outside the ecotone zone. It is difficult to follow the changes in the vegetation communities of these areas, because the main type of floras, i.e., warm-temperate and mesophilic, persisted throughout the entire Paleogene.

In the northern regions (Rarytkin Range, northwestern Kamchatka, Penzhina Bay), the transitional floras are found in the upper parts of the Umka, Ust'-Penzhina, and Irgirnin formations. These floras contain Platanaceae and the species close to *Trochodendroides* (against the background of the abundant *Cercidiphyllum*), as well as the ferns *Dennstaedtia*, *Osmunda*, *Ginkgo*, and *"Acer" arcticum*, with a large variety of modern genera of Amentiflorae, Rosaceae, and Aceraceae. Among them, one can recognize certain ancestors of some modern species. These floras are also distinctive because of a wide variety of conifers. The Rarytkin flora, besides *Sequoia affinis* and pines, also contains *Podocarpus*, *Taxus*, and *Taiwania* (Akhmet'ev, *et al.*, 1989).

In the southern regions of the Far East, the transition from the Eocene to Oligocene floras can be traced in the upper part of the coal-bearing lower Due Formation of Sakhalin and in the upper horizons of the Uglovaya Group in southern Primor'e, as well as in the age analogues of this group. The eastern Sikhote Alin' floral remnants contain a characteristic combination of small *Cocculus*-like leaves, which we attribute to the genus *Trochodendroides*, with the attendant frutification of *Nordenskioldia*, the small fruit of *Trochodendrocarpus*, and the leaves of *Platanus*, *Liquidambar*, *Zelkova*, *Ulmus*, *Alnus*, *Carpinus*, *Castanea*, *Acer*, Rosaceae, and others. Unlike the northern peri-Pacific floras, the transition floras of southern Primor'e have smaller

leaves. Both in the Eocene and in the Oligocene, the conifers spectrum was dominated by Taxodiaceae. The cooling in the Oligocene changed the composition of floras of the Lesnaya Formation in the Rarytkin Range. They begin containing *Picea, Larix, Alnaster, Betula* ex sect. *Fruticosae, Vaccinium*, and *Ledum*, though *Osmunda* and *Dennstaedtia* are still present.

The Zaisan basin is a classical region where the Eocene and Oligocene transitional layers show the remains of nearly all continental biota groups. Mixed floras with both subtropical and warm-temperate species are represented here by at least four sequential assemblages, and only the fifth Upper Oligocene Ashutas assemblage belongs to the typical Turgai flora without the taxons of Eocene subtropical origin (Akhmet'ev *et al.*, 1989). A sequence of transitional floras begins with that of the Upper Eocene, the so-called "swamp rhinoceros" flora, owing to the finds of *Zaisanamynodon borissovii Bel.* in the silty clays under the base of the flora-bearing layer. The "Eocene assemblage" of this flora is represented by Lauraceae (*Cinnamomum, Lindera*), Myrtaceae (*Eucalyptus*), xeromorphic poplars, and Magnolia. At this level, among the dominant species, there are the last *Hibiscus neuburgae* Iljinsk. The fine interbeds of the terminal "leaf" layers contain *Alnus, Ulmus, Zelkova*, and other deciduous plants. On the whole, the swamp rhinoceros flora indicates a humid climate, transitional from subtropical to warm-temperate environments.

The second flora of the sequence is confined to the lens lying on the contact between the Turanga and Tuzkabak formations that are assumed as the Eocene-Oligocene boundary. This interval of the section also includes the so-called Brontotheridae assemblages of vertebrate, remnants of the Late Eocene and the beginning of the Oligocene. The flora is xeromorphic, which agrees with the overlying red clay unit. The dominant plants are narrow-leaved poplars (*Populus kryshtofovichii, P. arnaudii*), Myricaceae (two species of sclerophyllous *Myrica*), and two species of *Eotrigonobalanus* (= *Dryophyllum*); also present are the small-leaved *Zelkova, Lindera*, and others, as well as temperate climate taxons, but only as ingredients. Both floras discussed above are free of conifers except rate exemplars of *Taxodium*.

The flora from the middle part of the Tuzkabah Formation is different. It occurs in that interval of the section where the Brontotheridae assemblage of vertebrates is already substituted by the Indricotheridae. The genera of the Turgai type make up 80% of the flora still retaining xeromorphism, which is much weaker than that of the preceding one. The species of *Taxodium, Comptonia, Populus arnaudii, Alnus neuburgae*, and *Cocculus schischkinii* dominate, and there are several species of *Quercus*. The Anacardiaceae are variable (*Rhus, Pistacia*, and *Astronium*). *Liquidambar, Platanus, Ailanthus*, and *Encommia* also occur, together with other thermophilic representatives of the warm-temperate deciduous flora, when *Populus arnaudii* was a migrant from the Upper Eocene.

The flora of the Buran Beds can be regarded as transitional between the Early and Late Oligocene assemblages. It precedes the Ashutas flora of the real Turgai type. Occurring in alluvial facies are the larger-leaved species of *Eucommia, Tilia, Alangium, Alnus, Ulmus*, and *Sassafras*. The remnants in limnetic facies are of smaller leaves, Salicaceae and Juglandaceae, which are dominant. *Populus arnaudii* remains present, though it appears for the last time in the Zaisan sequence. All Lower Oligocene floras of the Zaisan basin have xeromorphic features. The humidity of the climate, however, was ever changing, with two more dry periods: at the beginning and at the end of the Oligocene (the Solenovskii horizon). The red beds occur at the base and near the top of the Tuzkabak Formation (the Buran Beds).

Armenia

The Landzhar section in the Erevan-Ordubad depression was recovered by two drill-holes more than 500 m deep (Zaporozhets, 1989; Krasheninnikov *et al.*, 1989). In the lower marly part of the section, the entire sequence of planktonic foraminifers zones was identified (P15 - P17). The Oligocene beds are represented by marls (the bottom part of the section) overlain by clays and then by sandstones. They can be assigned to the *Globigerina tapuriensis* (P18), *G. sellii* (P19) zones, and probably also in part to the *G. ampliapertura* (P20) zone. The Upper Eocene and part of the Lower Oligocene sequences of the section (the basal layers of zone P18) show a high content of gymnosperm pollen, which is related to cooling started in the Upper Eocene (zone P16). The peak of the cooling is observed exactly between the P17 and P18 zones, assumed to be the Eocene-Oligocene boundary. Though the angiosperm pollen attains only 9%, being poor in spectrum, the generally subtropical pollen type indicates that the climate at that boundary remained subtropical.

The pollen-spectrum diagrams show irregular (counterphase) distribution of xerophilic and mesophilic components. The former are represented by the pollen of *Ephedra, Cedrus*, Cupressaceae, herbs, and small bushes (Chenopodiaceae, *Artemisia*); the latter constitute the spore part of the spectrum. The only cause of the compositional dynamics of palinospectra can be the alternation of humid and arid phases, because the proportion of the corresponding components totals 25% and more. The first of the six humid phases occurs in the P16 zone; the second, in the basal part of the P18 zone; the other four, within the undivided P19 and P20 zones.

As for the five arid phases, the first, second, and fourth were the longest. The first two phases correspond to the lower part of the P18 zone (*Ephedra* > 17%) and to the transition interval from the P18 to the P19 zone (the sum of pollen of the herbs and *Ephedra* reaches 20%); the last three phases correspond to the P19 and P20 zones, the fifth occurring just before the exsiccation of the sea basin.

Northern Ustyurt, the Aral region, Turgai

In northern Ustyurt, the studied flora was collected from the cores of the reference section recovered by hole 3006, drilled west of the Aktumsuk Cape in the Aral Sea, and from the type sections of the Chegan and Ashcheairyk formations in the chinks of northern Ustyurt. The data from hole 3006 shows that the boundary of notable cooling is distinctly evident in the Upper Eocene part of the section within the undivided *Globigerapsis tropicalis* and large *Globigerina* zones (P15 - P18). At that level, the pollen of Fagaceae, which was abundant in deposits of the Middle and the Early Upper Eocene, is substituted by gymnosperm pollen, reaching 90% of the spectrum (Krasheninnikov *et al.*, 1989).

The arid phase, so obvious in Armenia at the beginning of the Oligocene, is confirmed in this region as well (a higher content of *Cedrus* and *Ephedra* pollen).

The sections in the northern Ustyurt chinks show the most complete record of the transition interval history. The type section of the Chegan Formation near Sandal Mountain has both members: the lower clays and the upper siltyclays (the Turanga Beds). The Eocene-Oligocene boundary, as determined from dinoflagellate cysts, lies at the boundary between the *Wetzeliella clathrata angulosa* and *Phthanoperidinium amoenum* zones, 3 m below the top of the formation. This level coincides with the uppermost layer of the section composed of concretions bearing the Chegan type of mollusks. The dinocyst assemblage includes a group of species common for the *W. clathrata angulosa* Zone: *Deflandrea phosphoritica* (with subspecies), *D. heterophlycta, Thalassiphora pelagica, Areosphaeridium diktyoplokus, A. arcuatum, Cordosphaeridium*, and others. It should be noted however, that only a few representatives of the zonal index-species were observed. The diversity of organic-walled phytoplankton notably decreases in the upper Chegan member, being accompanied by the maritime index decline. The pollen assemblage of the Chegan Formation consists of two groups, which are most abundant: Pinaceae (mostly pines) and Fagaceae, with the dominating fine pollen of evergreen oaks *Quercus gracilis* and *Q. graciliformis* (40 - 60% of the angiosperm pollen spectrum). The pollen of Taxodiaceae, Juglandaceae, Moraceae, Hamamelidaceae, *Nyssa*, and *Rhus* is of minor importance. The Late Chegan period of cooling influenced the composition of the pollen assemblage. At precisely the Eocene-Oligocene boundary, the pollen of pines totals 90 - 95%, whereas that of Fagaceae is clearly reduced. This period of cooling, at first glance, seems to be discordant with the occurrence of thermophilic mollusks in the upper concretion beds of the formation. However, as the basin became shallower, even at lower temperatures the well warmed upper water layer could be favorable enough for the thermophilic fauna to survive.

The Lower Oligocene Ashcheairyk Formation conformably lies over the Chegan Formation but is lacking obvious hiatuses only in the axial part of the northern Ustyurt depression. The evidence of erosion of the underlying deposits can be found in fact everywhere at the contact of the two formations. The basal beds of the Ashcheairyk Formation belong to the *P. amoenum* Zone and have recorded events more clearly evident in the neighboring, more eastern regions of northern Aral and Turgai. These layers actually reflect the new state of the sea basin environments, i.e., lower salinity and temperature and a general lowering of the sea level to 60 - 80 m in amplitude. The latter is indicated by the depth of ancient valleys erosionally incised into the Chegan sediments and filled with the Kutanbulak Formation deposits in the northern Aral region. The degree of regression can be indirectly estimated by the area covered with retrogressive erosion marks produced after the sea level fall. The regression involved the entire Aral region and reached Turgai; it ceased only with the onset of the Ashcheairyk transgression.

The structure of the Ashcheairyk Formation is rhythmic, with three transgression-regression cycles. In regressive facies, the dinoflagellates practically disappear, being replaced by *Micrhystrydium, Pediastrum,* and *Hydropteris*. In transgressive facies, the abundance of dinoflagellates and their species diversity is restored. The composition of the dinofacies in the Ashcheairyk Formation changes within the section. The most freshwater group is represented by the succession of *Cyclonephelium-Adnatosphaeridium-Glaphyrocysta* and by *Phthanoperidinium*. The salt-water intervals contain *Deflandrea* and *Spiniferites*; the intermediate group consists of the Wetzeliellaceae. The cyprine-bearing beds are confined to the transgressive rhythms of the middle and upper unit and belong to the *Wetzeliella gochtii* zone. The pollen spectra of the formation are rich in Betulaceae and Juglandaceae, but conifers are dominant. The Pinaceae group becomes enriched in several species of Picea, Tsuga, and Abies. The Taxodiaceae to *Pinus* relation constantly changes, most likely reflecting the alternation of arid and humid climatic phases. In the Amentiflorae group, the pollen of *Carpinus* and *Carya* has a notable significance. Typical Oligocene species appear in this interval, i.e., *Juglans polyporata, Betula trigona*, and *Carya spackmannia.*

The transition from the Ashcheairyk Formation to the Solenovskii Horizon is gradual. In the Ashcheairyk Gully and west of it, a continuous bed of ferrugineous coarse-grain sandstones with fish remnants (vertebrae, shark's teeth) appears in the upper part of the Ashcheairyk Formation. The overlying sand-clay unit (10 m) still contains cyprines in its lower part, indicating that it still belongs to the Ashcheairyk Formation. There is another bed of ferrugineous sandstone ("the plate") at the top of this unit with numerous fragments of gnathic apparatuses, the vertebrae of fishes, shark's teeth, and gastropods. This bed and the underlying clays contain abundant valves of the brackish-water *Rzehakia cimlanica* and ostracods. The pollen assemblage of the Solenovskii Horizon resembles that of the Ashcheairyk Formation, though its spectra rapidly change, both within the section and laterally. The phytoplankton

assemblage indicates desalting of the basin, according to the appearance of *Ovoidites, Micrhystrydium, Pediastrum*, and *Cyclonephelium*, the last being very rare. The leaf flora of the upper Ashcheairyk Formation and the Solenovskii Horizon includes the broad-leaved plants typical of the Turgai type (*Alnus, Acer, Fagus*, and *Liquidambar*) and several subtropical plants inherited from the Eocene: *Myrica lignitum, Apocynophyllum* sp., and Lauraceae. The upper parts of the Ashcheairyk Formation and the Solenovskii Horizon can be well correlated with the upper beds of the Kutanbulak Formation and with the larger part of the Chilikta Formation.

The middle and upper units of the Chilikta Formation correspond to the level of the Solenovskii Horizon in the sections of the Bay of Perovskii in the Aral Sea. In the first unit, the loose ferrugineous sandstones are extremely rich in tooth remnants of the sharks *Odonthapsis, Rzehakia cimlanica*, and *Corbula sokolovii*. According to S.G. Zhilin, I.V. Vasil'ev, G.S. Rayushkina, and our data, the flora of the Kutanbulak Formation and of the Lower Chilikta Formation beds is of transitional type. Its composition combines subtropical and arctoboreal species. The flora of the upper beds of the Chilikta Formation begins to be similar to the typical Turgai flora. Both Lower Oligocene formations of the Aral Region are also traceable further eastward into Turgai, where they are known under the local names of the Urkimbai and Chelkar-Nura formations. The latter, in the eastern part of Turgai, is substituted by red deposits of the Saryin Formation. The underlying Urkimbai Formation wedges out toward the eastern side of the depression. In a typical section of Chelkar-Nura along the "Pin-Sai," where the first *Indricotherium transouralicum* was found, the Urkimbai Formation is represented by a rhythmical sequence of flat- and cross-bedded aleurolites (up to 15 m). In the lower part, the redeposited phytoplankton from the underlying Chegan Formation is combined with taxons typical of the *Phthanoperidinium amoenum* Zone, including the zonal species. The structure of the Chelkar-Nura Formation is more complicated but no less reliably dated. Like the Chilikta Formation, it has three main units (clay-sands - clay), persisting eastward into sections of the Saryin Formation. The middle sandy unit in the Chelkar-Nura contains remnants of large rodents, which were also found at the same level in Dongustau and in the Dulygalin Mosque sequences, along with remains of xerophytic flora containing *Rhus turomanica, Palibinia*, and *Cinnamomum*. Similar flora was observed in the loose Zhaman-Kaindy sandstones, which have the same stratigraphic position in the section of the Saryin Formation. The dinoflagellate assemblage with *Cyclonephelium* was recognized at the base of the upper clay unit of the Chelkar-Nura Formation in the Pin-Sai locality, thus providing data to tentatively correlate the unit with the Solenovskii Horizon. In Dongustau, large accumulations of shark's teeth were recorded at the same level. The *Indricotherium* assemblage of vertebrates in the axial part of the South Turgai depression were recovered either from the upper clay unit (in Pin-Sai, the bone-bearing lens lies nearer to its top) or from the underlying sand unit. This is the typical position of all large rhinoceroid burials in Turgai (Kyzyl-Kak, Mynyske-Suek, Saty, and other localities). The pollen spectrum in beds with *Indricotherium transouralicum* of the Pin-Sai locality contains up to 70% of Cheno- podiaceae, indicating the existence of salt-bearing marshes (or saltierra) at the end of the Early Oligocene in Turgai. At the same time, the plant remnants in the upper clay unit of the Mynyske-Suek section belong to the riparian facies, and all are of the Turgai flora type: *Strelitzia, Alnus ex gr. schmalhausenii, Populus balsamoides*, and Nymphaceae. A similar assemblage with riparian and interfluve plain plants is found in the basal layers of the coal-bearing Bollatam unit at the foot of Berdy Mountain. The unit conformably overlies the Saryin Formation and contains *Taxodium dubium, Alnus, Betula, Carpinus, Cyclocarya, Acer*, and others.

The changes in the paleogeographic environments in Turgai and in the Aral region were the direct result of oscillations of the Ashcheairyk Sea. They determined, or at least influenced the rhythm of climatic changes. The rising sea level swamped the coastal plains, partially destroyed vegetation, and expanded the area of halophytic assemblages. When the transgressions were at their culmination, the water discharge from the Turgai depression had to cease, which made the formerly dammed lake turn for a period into a sea basin. Since the beginning of regression in the Late Oligocene, along with exsiccation, the elevated areas were covered with polidominant conifer-broad-leaved mesophilic forests of the Turgai ecotype. The vast lowest part of the basin, now framed by the Zhilanchik arc, remained swamped and covered with Taxodiaceae and *Taxodiaceae-Alnus* forests in combination with abundant water plants. This vegetation type served as a good source of coal-forming material (the Bollatam-Zhilanchik coal basin).

As for the general scheme of the flora transformation in northwestern Kazakhstan at the time of the Eocene to Oligocene transition, it can be summarized as follows:

Upper Eocene	Lower Eocene	Upper Oligocene
"Paratropical" (subtropical) forests with dominant evergreens Lauraceae and Fagaceae	Savanna-like forests on interfluve plains: Taxodiaceae, Nyssaceae-Taxodiaceae and *Alnus, Salicaceae-Alnus* forests on the banks of water basins, Chenopodiaceae marshes on recently exsiccated coastal plains	Warm-temperate humid, conifer-broad-leaved forests of the Turgai ecotype

Western and Central Europe

In Western Europe, the "seam" of Late Paleogene transitional floras is diachronous. In the more northern regions (southern England, the Paris basin, Germany, Czechoslovakia), a notable ingress of boreal components started in the middle of the Early Oligocene (near the boundary of nannoplankton zones NP22-NP23). This time period corresponds to the Solenovskii Horizon in the eastern part of the Protoparatethys. This migration reached more southern regions at a later period. At any rate, the Tard clays (Hungary) with faunas of the Solenovskii Horizon in their upper part still contain paratropical flora. This flora started to disappear in the Kiscellian (the Chattian) and ended in the Egerian, at the very end of the Oligocene. In western Bohemia, the transitional floras are found in the Cheb and Sokolov basins and in the Dupov volcanic complex (Buzek *et al.*, 1990). The Upper Eocene floras of the Staro Sedlo Formation were still paratropical, bearing *Rhodomyrtophyllum* and *Steinhaurea*. The last occurrence of the *Normapollis* pollen was recognized there. The main components of the flora are Lauraceae and the evergreen Fagaceae. Two flora types were distinguished in the Lower Oligocene beds of the Dupov volcanogenic complex. The earlier one is transitional, being characterized by the Kuchlin flora type from diatomites near Bilin, which includes pines, the ferns *Acrostichum* and *Pteris*, palms, *Platanus neptunei*, and the evergreen Fagaceae. Most of the Late Eocene Lauraceae do not migrate into this flora stratum, but a small amount of the "arctoboreal" components is still present. The younger type is exemplified by the Kondratice flora of mixed mesophilic forests, with a high content of deciduous plants (*Ulmus*, birch). At the top of the Iosif Bed in the Novo Sedlo Formation, the thaphocoenosis includes the *Taxodiaceae-Alnus* assemblage valley forests.

In the Paris basin, the change from paratropical flora to the transitional type occurred almost exactly at the Oligocene-Eocene boundary, though as far back as the Late Eocene, the thermophilic elements were already being substituted by conifers and herbaceous plants, implying cooling and a transition to an arid climate. This is reflected in the composition of flora from the Pantin marls of the P17 zone. The transitional flora time coincides with the well-known jump in vertebrate evolution, the so-called "grande coupure," when these European Early Oligocene communities were supplemented by migrants from Asia. In the more northern regions of Europe, the change in the vertebrate assemblages (the boundary between 20 and 21 zones of Mein) almost exactly coincides with the time of expansion of the boreal floral migrants. This change took place in the Rupelian, at the boundary between the nannoplankton zones NP22-NP23 and more or less synchronously in southern England, the Paris basin, Rhein Graben, and Czechoslovakia.

North America

In the Late Eocene and Oligocene, as in Eurasia, there was a distinct transitional ecotone zone in the temperate latitudes of North America, where floras were most sensitive to climatic changes. The richest and best studied are localities in Yellowstone Park (the Florissant Formation) and westward into Oregon (the Clarno and John Day formations). The climate and depositional environments of the Florissant Formation, as indicated by the flora, were very similar to those of Zaisan. The Early Oligocene flora was xeromorphic and small-leaved, representing a combination of subtropical (including evergreens) and warm-temperate plants. The genera spectrum contains Cupressaceae, *Salix*, narrow-leaved *Populus*, various Anacardiaceae (*Rhus*, *Pistacia*, and *Astronium*), small-leaved Rosaceae (*Crataegus, Rosa*, and *Spiraea*), etc.

The floral succession in the Clarno and John Day formations (Manchester, 1990) was always regarded as a classic example of transition from the paratropical Eocene type to the warm-temperate Oligocene flora. One of the type floras of the Clarno Formation, that of the Nut Beds, records more than 150 fruit genera and 65 leaf genera, 60% of which are entire-margined. Most of them are evergreens, i.e., palms, *Cycas*, bananas, various vines (Menispermaceae, Icacinaceae, and Vitaceae), and arboreal plants (*Alangium, Mastixia, Nyssa*, and others).

Formerly, this flora was suggested to be from the terminal part of the Eocene, but new radiological datings of the flora-bearing ignimbrite unit reset the age at 43 - 47.4 Ma. Moreover, ignimbrites from the basal layers of the John Day Formation, lying 30 m higher up in the section, are still of the Late Eocene age (37 Ma). The other classic flora of the Clarno Formation is from the Cherry Creek locality and contains evergreens and deciduous species (*Ailanthus, Alnus, Platanus, Ulmus*, and others); it was also found to be older, possibly Middle Eocene, because the tuffs with the flora-bearing unit are 44.6 Ma old.

The floras in the basal layers of the John Day Formation (Bridge Creek, Erren Ranch, and other localities) already contain up to 70% of deciduous plants (*Metasequoia, Acer, Ulmus, Cercidiphyllum, Fagus, Quercus*, and others); they also have up to 10% of the extinct genera. The contrast between the Clarno and John Day floras from the monofacies sediments is striking. They have, however, the 10 - 12 Ma time difference, and it is as yet difficult to state at what stage the most spectacular changes in flora took place. We are certain only that the cooling started as far back as the Late Eocene.

A study of the transitional interval from the Eocene to the Oligocene in different regions shows that the sequential transformations of ecosystems in the entire Boreal and in the northern part of the Tethyan areas had the same trend and similar scenarios.

Climatic cooling and regression of the epicontinental seas from higher latitudes of both continents in the

Northern Hemisphere, which started in the middle of the Late Oligocene, gave an impulse to the breakdown of stable biocoenoses. These events caused a principal reformation in the zonal types of forest vegetation, a change of forest-forming species, and a compositional alteration of intrazonal communities, particularly in the ecotone zone. The recurrence of humid and arid phas- es in climate in the Early Oligocene accelerated the destruction of the old and the establishment of the new zonal types of vegetation. In the Late Oligocene, another impulse of cooling finally defined the warm-temperate mesophilic forest vegetation of the Turgai ecotype in the middle latitudes of Eurasia and North America.

CONCLUSIONS

1. The most obvious indications of cooling appeared mainly in the middle of the Late Eocene, yet in some provinces it took place nearly at the time of Middle-Late Eocene boundary. The cooling was not sudden, but progressing to the pessimal peak at the Eocene-Oligocene boundary. In the first half of the Early Oligocene, there was a widespread slight rise of mean annual temperatures, and these conditions lasted until the Late Oligocene with small variations. At the end of the Oligocene, the mean annual temperatures lowered again and reached absolute minimal values for the Paleogene.

2. As evidenced by the compositional change of the Lower Oligocene floras from the ecotone zone, the transformations in vegetation cover were caused not so much by the fall of the mean annual temperatures, but rather by the alternation of arid and humid phases. The data from Armenia indicates that in the Early Oligocene alone, this alternation of phases occurred five times. Actually, these fluctuations were fatal to many plants widely distributed in the first half of the Paleogene (the species *Dryophyllum, Trochodendroides, Hibiscus neuburgae*, narrow-leaved poplars of the type *Populus arnaudii* and *P. celastrophylla*, *"Acer" arcticum*, and others).

3. In the ecotone zone, the flora changes continued for no less than 5 Ma. The seam of subtropical floras was substituted first by the assemblages of transitional plants and then by mesophilic temperate-warm plants. The transitional floras were characterized by combination, repeatedly changing in time, complicated by more temperate deciduous mesophilic components with subtropical plants, characterized by the dominating evergreen Lauraceae, Fagaceae, and *Populus*. The latter proved to be more resistant in riparian formations, which they abandoned only in the beginning of the Late Oligocene.

4. The leading factor, which affected the changes in ecosystems and the whole history of climatic processes, was the exsiccation of the meridional sea gulfs (Turgai and Mississippi) and of much larger aquatories of epicontinental seas (western Siberia) representing parts of the warm-water ocean areas.

5. Despite the beginning of cooling, the climate in the Late Eocene within the ecotone zone remained subtropical and humid. Only in Kazakhstan, particularly in its eastern part, were conditions transitional from subtropical to warm-temperate. The same kind of climate was in southern Primor'e; but in the other parts of the Far East and northeastern Asia, the climate was warm-temperate. In the first half of the Oligocene, the climate remained subtropical in central and southern Europe and all over the Crimea-Caucasus area, where only in the eastern part was the climate periodically arid. In the northern Aral, Turgai and Zaisan regions, the climate of the Early Oligocene was periodically arid, probably seasonal, similarly transitional from subtropical to temperate-warm; in the Pacific region, the climate was humid and temperate-warm. The differences in climatic conditions of various regions in the Early Oligocene defined the individual features of ecological systems in accordance with their rank. This ultimately was reflected both in the scale and in the characteristic features of ecological changes.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 93 - 99.
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Stratigraphy of the Jurassic to Lower Cretaceous Cherts and Terrigenous Deposits of Primor'e (Gorbusha Strata)

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Manuscript received February 20, 1992

The most representative Mesozoic section of Eastern Primor'e is exposed in the Dal'negorsk area. It is composed of cherts and terrigenous deposits of the so-called Gorbusha Strata. All the three series of the Triassic, the Lower and Upper Jurassic, and the Lower Cretaceous have been identified in these thin (300 - 400 m) unit. However, stratigraphic subdivision of these deposits was rather poorly substantiated until now. The Triassic part of the sequence is distinctly subdivided into zones according to conodonts and radiolarians (Bragin, 1988; Bragin *et al.*, 1988), but Jurassic and Lower Cretaceous deposits containing radiolarians have not been subdivided to such details, and their ages have been defined to within at an epoch rank. This was caused by the fact that the radiolarians from these strata were rather poorly studied. This paper attempts to make up this deficiency. The structure of the Gorbusha Strata has been described by Bragin *et al.* (1988), who have shown that four tectonic slices can be identified in the type section of the sequence along the Rudnaya River.

In the section of the second tectonic slice (Krasilov and Parnyakov, 1984) (bed 2/17), the Upper Triassic gray, light-gray, and lilac-gray platy glassy chert, with late Carnian to middle Norian *Triassocampe nova* Yao and other radiolarians, is overlain without a visible break by Lower Jurassic deposits (Bragin *et al.*, 1988). A bed-by-bed description of these deposits (from the base upsection) is given below.

J_1p? Bed 2/18. Light-gray and gray massive, less frequently, platy chert with the radiolarians *Parahsuum simplum* Yao, *P. longiconicum* Sashida, *P.* cf. *ovale* Hori et Yao, *Jacus* cf. *sandspinensis* Pessagno, Whalen et Yeh, *Praeconocaryomma immodica* Pessagno, and *Trillus* sp. of the Early Jurassic, most likely of Pliensbachian age. Thickness is 5 m.

J_1p_2 - t_1. Bed 2/19. Platy glassy chert, from dark-gray to black in the lower part of the bed and light-gray upsection, sometimes with small inclusions of carbonaceous detritus. The late Pliensbachian to early Toarcian radioalarians *Parahsuum longiconicum* Sashida, *P. simplum* Yao, *P. kanyoense* Sashida, *Parvicingula* (?) aff. *gigantocornis* Kishida et Hisada, *Jacus* cf. *sandspinensis* Pessagno, Whalen et Yeh, *Praeconocaryomma immodica* Pessagno, *Trillus* aff. *elkhornensis* Pessagno et Blome, *Triactoma* sp., *Hagiastrum sp.*, and *Hsuum* sp. were found in the lower part of the bed (up to 4.5 m from the base) (Table 1). The middle Toarcian radiolarians *Crubus wilsonensis* Carter, *Hsuum* cf. *rosebundense* Pessagno et Whalen, *H.* sp. *B* (Carter *et al.*, 1988), *Paronaella* cf. *variabilis* Carter, *Praeconocaryomma immodica* Pessagno, and *Pseudocrucella* sp. were found upsection (Table 2). The fragments of redeposited Triassic conodonts were found in the lower part of the bed. Thickness is 8 m.

J_3cl - t_1. Bed 2/20. Gray-green clayey tuffaceous chert. The lower boundary of the bed is very sharp, but without any traces of erosion or nonconformity. The Callovian to Oxfordian radiolarians *Eucyrtidiellum ptyctum* (Riedel et Sanfilippo), *Gueuxella nudata* (Kocher), *Hsuum* ex gr. *maxwelli* Pessagno, *Protunuma* cf. *ochiensis* Matsuoka, *Stichocapsa robusta* Matsuoka, *Ristola dhimenaensis* (Baumgartner), *Stylocapsa oblongula* Kocher, and *Tricolocapsa pentagona* Aita were found in the lower part (2 m) of the bed. Upsection, abundant Kimmeridgian to early Tithonian radiolarians were found: *Archaeodictyomitra apiara* (Rüst), *Dictyomitrella* (?) *kanoensis* Mizutani et Kido, *Emiluvia orea* Baumgartner, *Mirifusus mediodilatatus* (Rüst), *Podocapsa amphitreptera* Foreman, *Ristola altissima* (Rüst), *Spongocapsula* aff. *perampla* Pessagno, and *Tetraditryma* aff. *pseudoplena* Baumgartner. Thickness is 7 m.

J_3t_2 - K_1v. Bed 2/21. Flyschlike alternation of gray-green chert (very clayey in the lower part of the bed) with dark-gray and black thin-platy claystone and siltstone, and gray fine- to medium-grained sandstone, massive and thick-platy, graded, sometimes cross-bedded. The thickness of the interbeds of rudaceous rocks increases upsection, and such rocks prevail in the upper part of the bed. The middle Tithonian to Valanginian radiolarians *Alievium helenae* Schaaf, *Ditrabs sansalvadorensis* (Pessagno), *Mirifusus mediodilatatus* (Rüst), *Pantanellium berriasianum* Baumgartner, *Pseudodictyomitra carpatica* (Lozynyak), *Ristola altissima* (Rüst), and *R. cosmoconica* (Foreman) were recovered from clayey chert and claystone. The redeposited Triassic radiolarians *Pseudostylosphaera* cf. *japonica* (Nakaseko et Nishimura), *Triassocampe* sp., and *Yeharaia* sp. were found in the sandstone of the upper part of the bed. Thickness is 35 m.

K_1. Bed 2/22. Gray massive fine- and medium-grained polymictic sandstone. Thickness is 35 m.

Analogues of Beds 2/20 and 2/21 were also found in the first tectonic slice of the type section of the Gorbusha Strata (Fig. 1), where they compose isolated tectonic blocks. One of these blocks consists of the following (from the base upsection):

$J_3km - t_1$. 1. Dark gray-green clayey tuffaceous chert containing the Kimmeridgian to early Tithonian radiolarians *Podocapsa amphitreptera* Foreman, *Tetraditryma corallitosensis* Pessagno, *Spongocapsula sp.*, and others. The visible thickness is 10 m.

Tectonic contact.

$J_3km - t_1$. 2. Similar chert with the Kimmeridgian to early Tithonian radiolarians *Archaeodictyomitra apiara* (Rüst), *Cinguloturris carpatica* Dumitrica, *Emiluvia orea* Baumgartner, *Mirifusus mediodilatatus* (Rüst), *Podocapsa amphitreptera* Foreman, *Ristola altissima* (Rüst), *Spongocapsula perampla* Pessagno, *Triactoma tithonianum* (Rüst), *Tritrabs ewingi* (Pessagno), and others. Thickness is 7 m.

$J_3t_2 - K_1v$. 3. Flyschlike alternation of claystone, siltstone, and sandstone containing the middle Tithonian to Valanginian radiolarians *Alievium helenae* Schaaf, *Ditrabs sansalvadorensis* (Pessagno), *Pantanellium berriasianum* Baumgartner, *Podocapsa amphitreptera* Foreman, and others. Thickness is 6 m.

Tectonic contact.

$T_2l_2 - T_3k$. 4. Light greenish-gray platy chert with the Triassic (late Ladinian to Carnian) radiolarians *Plafkerium cochleatum* (Nakaseko et Nishimura), *Sarla dispiralis* Bragin, and *Triassocampe* ex gr. *deweveri* (Nakaseko et Nishimura). Thickness is 5 m.

The lower beds of the Upper Jurassic are represented more completely in the area of the Sadovyi Settlement. There, the sequence consists of the following (from the base upsection):

T_3n_{2-3}. 1. Dark-gray massive or, occasionally, thick-platy chert with the middle to late Norian conodonts *Epigondolella abneptis* (Huckriede) and *E. postera* (*Kozur*). Thickness is 10 - 12 m.

Tectonic contact.

$J_3cl - t_1$. 2. Gray and lilac-gray thin-platy chert, in the upper part of the bed, clayey, dark-gray. The Callovian to Oxfordian radiolarians *Eucyrtidiellum ptyctum* (Riedel et Sanfilippo), *Protunuma ochiensis* Matsuoka, *Ristola dhimenaensis* (Baumgartner), *Stylocapsa himedaruma* Aita, *S. naradaniensis* Matsuoka, *S. robusta* Matsuoka, *Stylocapsa oblongula* Kocher, and *Theocapsa pentagona* Aita were found in the lower 15 m of the bed. The upper 5 m has been characterized by a Kimmeridgian to early Tithonian radiolarian assemblage: *Archaeodictyomitra apiara* (Rüst), *Mirifusus mediodilatatus* (Rüst), *Podocapsa amphitreptera* Foreman, and *Ristola altissima* (Rüst). Total thickness is 20 m.

$J_3t_2 - K_1v$. 3. Flyschoid alternation of gray clayey tuffaceous chert, black siltstone and claystone, and gray fine-grained sandstone. The radiolarians found there, *Alievium helenae* Schaaf, and *Ditrabs sansalvadorensis* (Pessagno), are of middle Tithonian to Valanginian in age. Thickness is 15 m.

The overlying gray medium-grained sandstone is, most likely, of the Early Cretaceous age.

An analysis of the above sections makes it possible to outline several successive members in the Jurassic to Lower Cretaceous Gorbusha Strata (from the base upsection):

	Thickness, m
1. Cherts of Pliensbachian-Toarcian age	13
Stratigraphic gap	
2. Cherts of Callovian-Oxfordian age	2 - 15
3. Clayey cherts of Kimmeridgian to early Tithonian age	5 - 7
4. Terrigenous flyschoids of middle Tithonian to Valanginian age	up to 25
5. Sandstones of Early Cretaceous(?) age	35 - 200

ANALYSIS OF RADIOLARIAN ASSEMBLAGES

Taking into account the distribution of the radiolarian assemblages (Table 1), the sequence under consideration can be subdivided (from the base upward) as follows:

1. Beds with *Parahsuum longiconicum-P. kanyoense*. Upper Pliensbachian to lower Toarcian.

The lower boundary of the beds is placed according to the first appearance of *P. longiconicum*, *P.* cf. *ovale*, and *Jacus* cf. *sandspinensis*. *P. kanyoense* and *Parvicingula gigantocornis* occur in the upper part of the beds. The transition species are represented by *Parahsuum simplum* and *Praeconocaryomma immodica*. The beds may be correlated with the *Parahsuum takarazawaensis* Zone of the upper Pliensbachian to the lower Toarcian from central Japan (Sashida, 1988) according to the common species *P. longiconicum* and *P. kanyoense*, whose stratigraphic range lies within this zone. The time range of the beds with *P. longiconicum-P. kanyoense* can be confirmed, as well, by the presence of *Jacus* cf. *sandspinensis*, described from the Pliensbachian of British Columbia (Pessagno *et al.*, 1986); *Parahsuum simplum*, known in the Pliensbachian of Japan (Hori and Yao, 1988; Sashida, 1988); and *Praeconocaryomma immodica* distributed in the upper Pliensbachian-Toarcian of western North America (Carter *et al.*, 1988).

In the sequence of the Gorbusha Strata, the beds with *P. longiconicum-P. kanyoense* overlie the Upper Triassic rocks with a hiatus, because the Hettangian and Sinemurian fossils are completely absent. Because of this fact, the exact stratigraphic position of the lower boundary of the beds cannot be determined.

2. Beds with *Crubus wilsonensis-Hsuum* cf. *rosebundense*. The middle Toarcian.

The lower boundary of the beds is defined according to the appearance of *C. wilsonensis*, *H.* cf. *rosebundense*, and *Paronaella* cf. *variabilis*, described from the middle

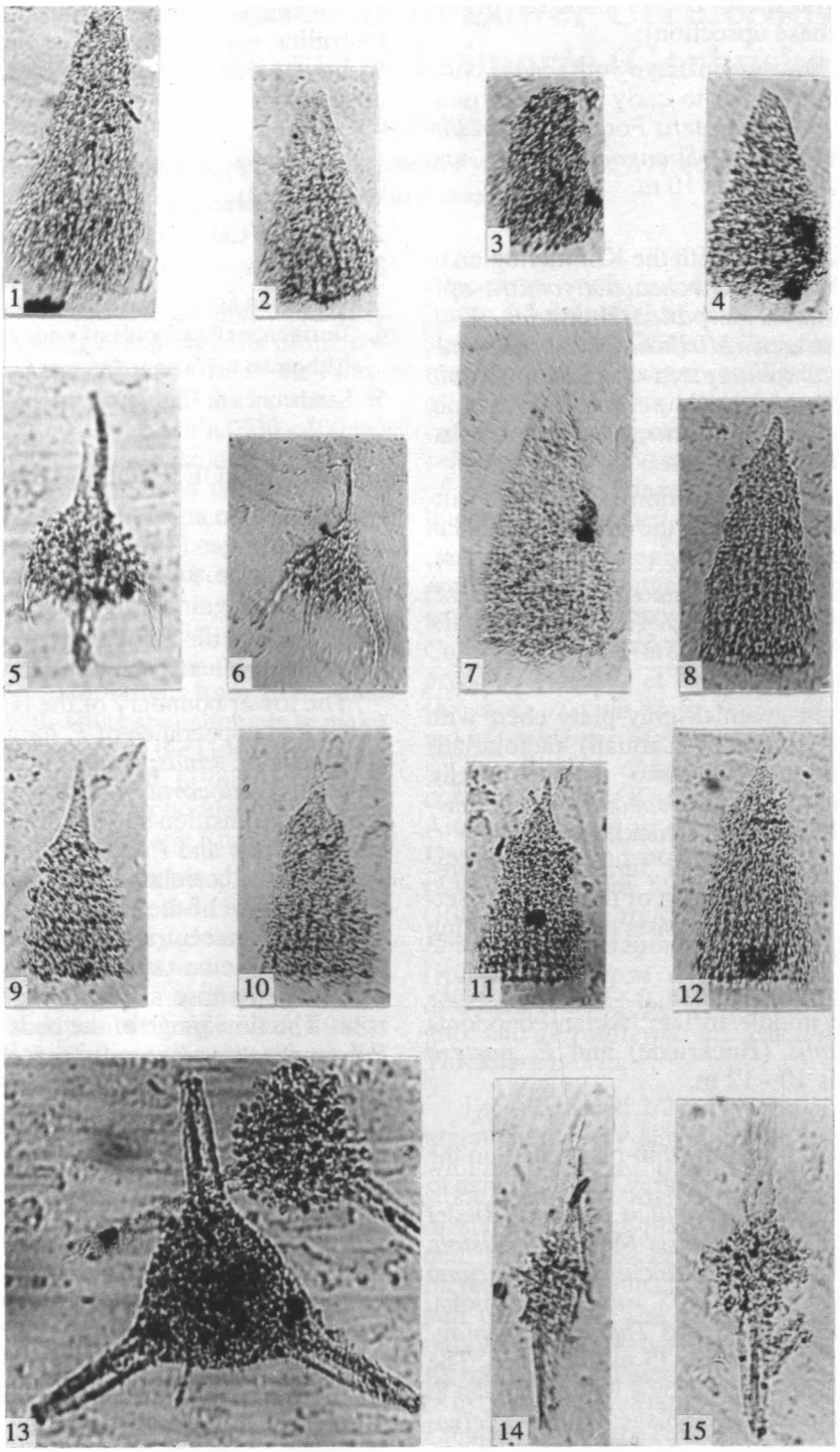

Plate I. *Radiolarians from the Pliensbachian to early Toarcian deposits.*
1, 2 – Parahsuum simplum Yao; 3 – P. cf. ovale Hori et Yao; 4 – P. sp.; 5, 6 – Jacus cf. sandspinensis Pessagno, Whalen et Yeh; 7, 8 – Parahsuum longiconicum Sashida; 9 – Parvicingula aff. gigantocornis Kishida et Hisada, 10 - 12 – Parahsuum kanyoense Sashida; 13 – Triactoma sp.; 14, 15 – Trillus aff. elkhornensis Pessagno et Blome. Magnification 120 (all forms).

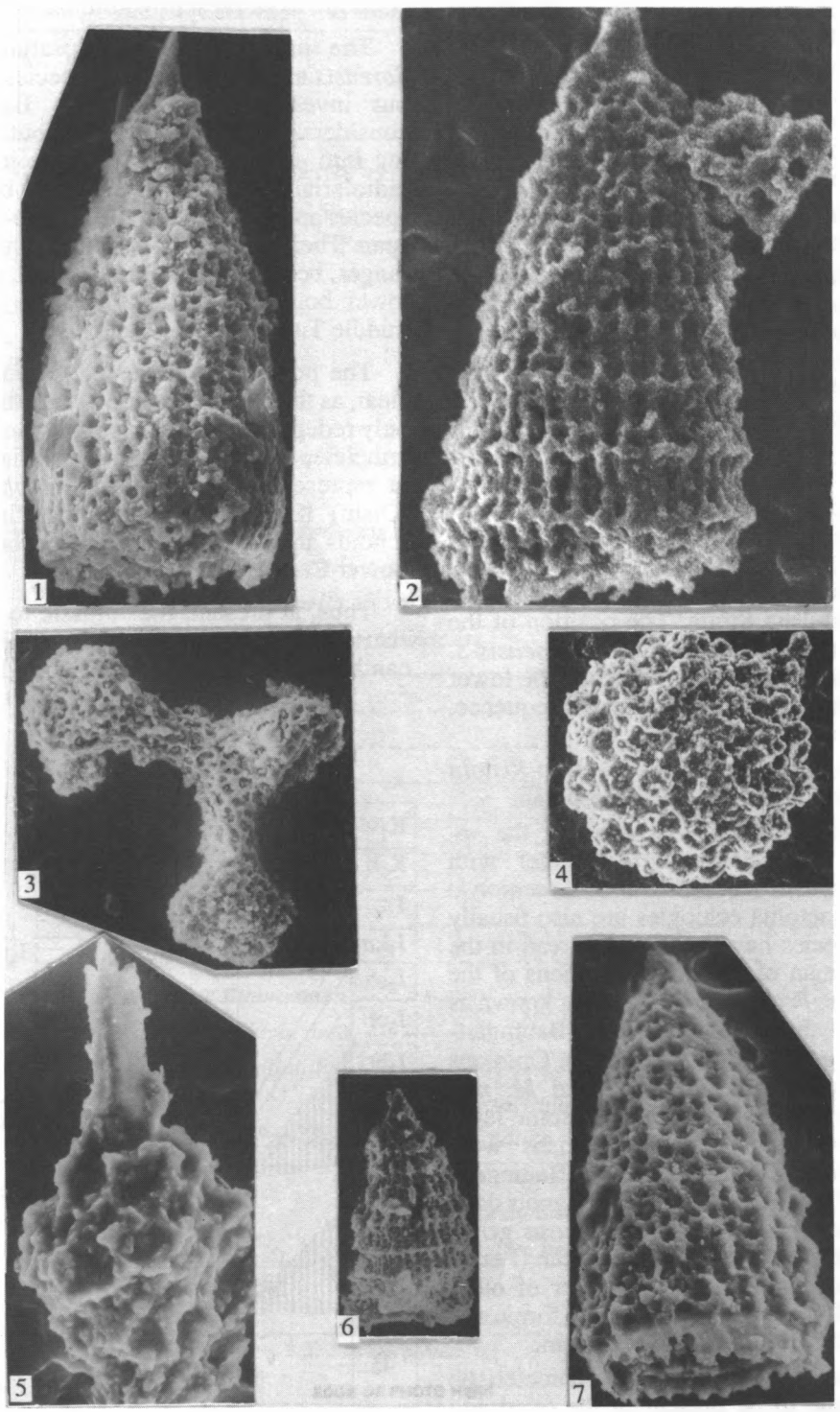

Plate II. *Radiolarians from the middle Toarcian deposits.*
1 – Hsuum cf. rosebundense Pessagno et Whalen. Magnification 200; 2 – Crubus wilsonensis Carter. Magnification 200; 3 – Paronaella cf. variabilis Carter. Magnification 200; 4 – Praeconocaryomma immodica Pessagno. Magnification 120; 5 – Trillus aff. elkhornensis Pessagno et Blome. Magnification 250; 6 – Grubus wilsonensis Carter, Magnification 75; 7 – Hsuum sp. Magnification 200.

Toarcian of western North America (Carter *et al.*, 1988). *Praeconocaryomma immodica* is a transition species. The beds correspond to the middle Toarcian.

Their upper boundary cannot be precisely defined, as the beds are unconformably overlain by the Upper Jurassic deposits. The Middle Jurassic is completely missing from the sequence. Minute study of the radiolarians showed that the supposedly Middle Jurassic deposits, identified formerly in the Dal'negorsk section (Bragin *et al.*, 1988), correspond in fact to the above-described Toarcian beds.

3. The beds with *Ristola dhimenaensis-Stylocapsa oblongula*. The Callovian-Oxfordian.

The lower boundary of the beds is drawn according to the appearance of *R. dhimenaensis*, *Hsuum* ex gr. *maxwelli*, and *Protunuma ochiensis*. Characteristically, there are such species as *Eucyrtidiellum ptyctum* and *Gueuxella nudata*. All these species have been widespread in the Callovian-Oxfordian of the Tethys and the tropical regions of the Pacific (Baumgartner, 1984; Aita, 1987).

At present, the Callovian and Oxfordian are subdivided according to radiolarians in great detail (Baumgartner, 1984; Aita, 1987). Unfortunately, the obtained data is insufficient to apply similar subdivision to the sequence of the Gorbusha Strata. The position of the lower boundary of the beds with *R. dhimenaensis-S. oblongula* remains so far uncertain. Perhaps the lower part of the Upper Jurassic is missing from the sequence, as with the Middle Jurassic.

4. The beds with *Podocapsa amphitreptera-Ristola altissima*. The Kimmeridgian to lower Tithonian.

The lower boundary of the beds is set at the appearance level of *P. amphitreptera*, together with *R. altissima* and *Mirifusus mediodilatatus*. *Acaeniotyle diaphrogona* and Triactoma echioides are also usually present. All these species have been widespread in the Kimmeridgian-Tithonian of the tropical regions of the Tethys and Pacific, the *P. amphitreptera* being known as appearing only in the Kimmeridgian (Baumgartner, 1984; Pessagno *et al.*, 1984; Aita, 1987). Opinions about the time ranges of *R. altissima* and *M. mediodilatatus* are contradictory. European and some Japanese authors believe that these nassellarians were already present in the Callovian-Oxfordian (Baumgartner, 1984; Aita, 1987), but according to Pessagno's data, *R. altissima* and representatives of the *Mirifusus* genus appeared for the first time in the Kimmeridgian (Pessagno *et al.*, 1984). A relatively small number of other groups of fossils in all the sections of both California and the Tethys hampers the solution of this problem.

It should be noted that radiolarians characteristic of these beds continue to be present in the overlying deposits as well, but those beds can be identified as an independent unit according to the first appearance of new taxa in them. Such practice is used in radiolarian zonal subdivision (Baumgartner, 1984; Aita, 1987).

5. The beds with *Ditrabs sansalvadorensis-Ristola cosmoconica*. The middle Tithonian to Valanginian.

The lower boundary is distinguished according to the appearance of *D. sansalvadorensis*, *R. cosmoconica*, *Alievium helenae*, and *Sethocapsa pseudoterculus*. *Ristola altissima* and *Mirifusus mediodilatatus* continue to be present within the beds.

The timing of the first appearance of *D. sansalvadorensis* and accompanying species is defined by various investigators. Thus, P.O. Baumgartner (1984) considers it to be late Tithonian, but Y. Aita (1987), taking into account the data on the joint distribution of radiolarians and nannoplankton, believes that these species appeared within the middle-late Tithonian time span. These different opinions and the radiolarian time ranges, being insufficiently studied, permit defining the lower boundary of these beds as approximately the middle Tithonian.

The position of their upper boundary is even less clear, as they are overlain by sandstone, which contains only redeposited radiolarians and no other fossils. Nevertheless, the finds of the Berriasian-Valanginian flora representatives, such as *Alsophylites nipponensis* (Dishi) Krassilov, indicate that the terrigenous flyschoids and sandstones of the Gorbusha Strata include Lower Cretaceous deposits.

Thus, at present, the Jurassic to Lower Cretaceous cherts and terrigenous deposits of the Gorbusha Strata can be subdivided, according to radiolarians, into fos-

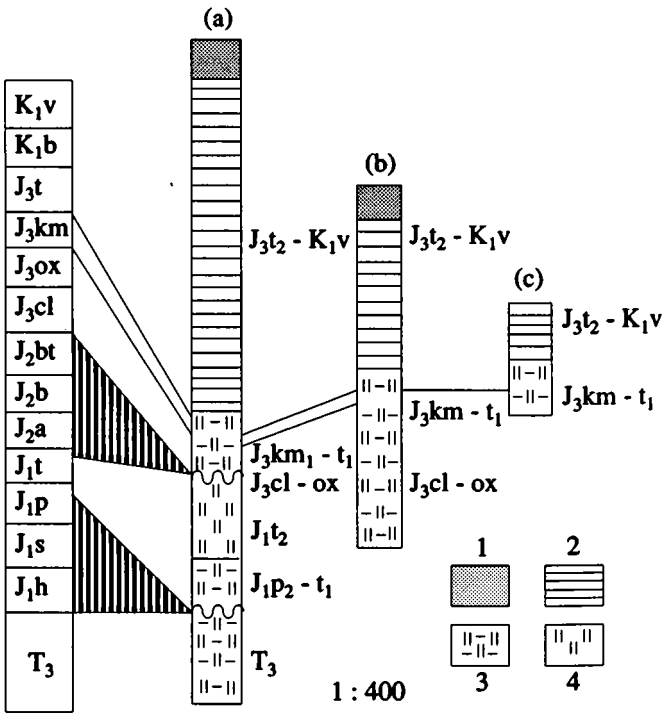

Fig. 1. *Sections of the Gorbusha Strata and their correlation. a – Reference section, tectonic slice II; b – Section near the Sadovyi Settlement; c – Reference section, tectonic slice I. 1 – sandstone; 2 – siltstone and claystone, 3 – clayey chert, 4 – chert.*

Table 1. Subdivision schematics of the Jurassic to Lower Cretaceous Gorbusha Formation according to radiolarians

Triassic	Jurassic					Cretaceous	System
Upper	Lower		Upper			Lower	Series
Norian	Pliensbachian	Toarcian	Callovian-Oxfordian	Kimmeridgian	Tithonian	Berriasian	Stage
	P. l. - P. k	*C. w. - H.* cf. *r.*	*R. d. - S. o.*	*P. a. - R. a.*	*D. s. - R. c.*		Beds with fauna
Cherts	Cherts 13 m		Cherts 2 - 15 m	Clayey cherts 5 - 7 m	Flyschoids 35 m		

Parahsuum longiconicum
Jacus cf. *sandspinensis*
Parahsuum simplum
Parahsuum cf. *ovale*
Trillus aff. *elkhornensis*
Praeconocaryomma immodica
Parahsuum kanyoense
Parvicingula aff. *gigantocornis*
Crubus wilsonensis
Paronaella cf. *variabilis*
Hsuum cf. *rosebundense*
Ristola dhimenaensis
Hsuum ex gr. *maxwelli*
Stylocapsa oblongula
Gueuxella nudata
Eucyrtidiellum ptyctum
Podocapsa amphitreptera
Acaeniotyle diaphrogona
Ristola altissima
Triactoma echioides
Triactoma tithonianum
Emiluvia orea
Mirifusus mediodilatatus
Ditrabs sansalvadorensis
Pseudodictyomitra carpatica
Ristola cosmoconica
Sethocapsa pseudouterculus
Alievium helenae

siliferous beds that can be reliably traced in different sections. The revealed radiolarian assemblages are typical of open marine and oceanic paleobasins and have a pronounced tropical affinity, which makes it possible to correlate with enough certainty these biostratigraphic subdivisions with the zones of the various Jurassic-Cretaceous scales compiled for the Mediterranean-Alpine region and the circum-Pacific belt. The results of the executed investigations have finally proved that there are significant hiatuses within the Gorbusha Strata, comprising the Hettangian and Sinemurian stages and the entire Middle Jurassic Series.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 102 - 104.
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Evolution of the biosphere

A More Precise Definition of the Commencement of the Lange Ranch Eustatic Event at the Cambrian-Ordovician Boundary

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Manuscript received March 25, 1992

It is known that the most appreciable rearrangement in the associations of the faunal groups such as trilobites and conodonts, corresponding to the base of the conodont *Cordylodus proavus* Zone (Chugaeva *et al.*, 1989; Dubinina, 1990), is proposed as the Cambrian-Ordovician boundary. Many researchers correlate faunal changes at this boundary with a world-wide eustatic event, which was called the Lange Ranch Event[1] by J. Miller (1984).

The Lange Ranch Event is most clearly manifested in the Lange Ranch section of central Texas. The commencement of this event (or the Acerocare Event (Erdtmann, 1986)) has also been recorded by J. Miller and B. Erdtmann (Erdtmann and Miller, 1981) at the base of the *Cordylodus proavus* Zone.

The synchroneity of this eustatic phenomenon is recorded in the majority of sections on different continents, but it is especially clear in the shallow-marine sections of platforms (Miller, 1984). Such an event ought to be regarded as glacial eustatic event according to both J. Miller and other researchers' point of view (Erdtmann, 1982; Fortey, 1984; Miller, 1984). The Lange Ranch Event affects all paleobasins, but it has specific features in every concrete setting. It is also necessary to take into account that this event was studied differently in the various regions.

Two disconformities, where short-term breaks in sedimentation occured, were established in the shallow-marine carbonate sequences of the *C. proavus* Zone in the western United States (western Utah, eastern Nevada, central Texas). The first disconformity is observed directly at the base of the *C. proavus* Zone (or at the base of the trilobite *Corbinia apopsis* Zone). The second one is observed somewhat stratigraphically higher, at the top of the *C. apopsis* Zone corresponding to the base of the Canadian Series. The Lange Ranch Event began with sharp and short-term sea level changes (the first phase of event), which were replaced by a relatively calm period of a gradually rising sea level (the second phase of event).

Accordingly, the oolitic and detrital limestones that prevailed during the initial stage were replaced by the thick and rather monotonous strata of stromatolitic and dolomitic limestones. By the *Cordylodus intermedius* time, the Lange Ranch Event was concluded (Miller, 1984), the sea level stabilized, and the low-energy sedimentation regime settled.

In the relatively deep-water paleoenvironments of the continental slope at the western margin of the North American continent, the Lange Ranch Event was displayed in the changes in the rhythm and velocity of sedimentation. These were pronounced, foremost, in different proportions of coarse- and finegrain material, which composes breccias and thin turbidites (Cook and Taylor, 1977; Cook and Egbert, 1981). The changes in the sedimentation regime of the continental slope were accompanied by similar changes in a shallow-water setting of adjacent platform margin. Thus, before the Lange Ranch Event, the Trempealeauan deposits in the Tybo Canyon (central Nevada) were represented by an alternation of thin turbidities and breccia interbeds, comparable in thickness with synchronous sediments of the platform margin (the Notch Peak Formation of western Utah). The accumulation of thin, shallow-water carbonates during the early interval of the event in western Utah was alternated with short-term breaks in sedimentation (Miller, 1984); thick strata of prevailed breccia with subordinate interbeds of thin turbidities were formed at that time in central Nevada. These deposits are the result of the breaking up and redeposition of material from the closest shallow-water areas. The thick stromatolite strata of the second phase of the event in western Utah and eastern Nevada are of the same age as the massive clastic deposits in central Nevada, related to subaqueous slides and collapses. Finally, the accumulation of the fine-grained House Limestone in western Utah and the Hot Creek calcareous mudstone with rare breccia interbeds in central Nevada indicate the stabilization of both the sea level and the sedimentation processes at the end of the Lange Ranch Event.

[1] The Black Mountain (Miller, 1984), or Peltocare (Erdtmann, 1986) Eustatic Event and the Ceratopyge Event (Erdtmann, 1986) preceded by the Lange Ranch Event are not considered in this paper.

It is interesting to compare the deposits of central Nevada and western Newfoundland, which were formed on the western and eastern continental slopes of North America, respectively. Regimes of sedimentation on these slopes were different during the Lange Ranch Event: quick accumulation of thick alternating breccia and thin turbidite occurred in the first case (Cook *et al.*, 1989a); intense erosional processes, bringing to "the loss" of contemporaneous layers, occurred in the second case (Stevens and Stouge, 1982). The deposits of the larger parts of the *Cordylodus primitivus* and *C. proavus* zones in the sections of western Newfoundland (particularly in the proximal and central parts of the basin) are absent as a result of erosion by debris flows, accompanied by the accumulation of megabreccias.

Another example of the Lange Ranch Event manifestation can be observed in Scandinavian sections. Many of them are deprived of sediments of the *Acerocare* Zone and the Tremadocian beds disconformably overlie older beds with *Peltura* (Erdtmann, 1982).

The more complex character of the eustatic sea level fluctuations near the Cambrian-Ordovician boundary has been registered in the shallow-water sediments of the Baltic-Ladoga glint (Popov *et al.*, 1989). Meanwhile, the maximum of the Lange Ranch regression in these sections also coincided with the base of the *Cordylodus proavus* Zone.

The regarded eustatic event in the Kazakhstan sections (Batyrbay and Sarykum) was studied in detail by V.G. Zhemchuzhnikov (1987). The first short-term shallowing of the basin up to the intertidal environment in the Batyrbay section has been found at the level of 107 - 107.5 m; the second shallowing has been revealed at 126.5 m, which corresponds, perhaps, to the repeated regression found by Miller in the shallow-water carbonate sections of the North American platform at the beginning of the Lange Ranch Event, not at the base, but at the top of the trilobite *Corbinia apopsis* Zone.

A single manifestation of the Lange Ranch Eustatic Event was recorded in siliceous-carbonate facies of the northwestern Balkhash area (the Sarykum section) (Zhemchuzhnikov, 1987).

The goal of this paper is to establish more accurately the commencement of the Lange Ranch Event; therefore, attention was paid to the Batyrbay section of the Malyi Karatau mountains. Recent minute sedimentological research by H. Cook and V.G. Zhemchuzhnikov (Cook *et al.*, 1989b, 1991), showed that in the Early Paleozonic, Malyi Karatau represented a carbonate seamount (called Aisha-Bibi), isolated from the continent. The Batyrbay and adjacent Kyrshabakty sections were situated in the northeastern margin of this seamount. Obviously, the relatively low intensity of erosional processes on the slopes of the Aisha-Bibi Seamount (as distinguished from those on continental slopes) has predetermined the preservation of sediments of the Lange Ranch Event's initial regressive phase in the Batyrbay section.

Let us look at the interval of 102 - 109 m of the Batyrbay section (Zhemchuzhnikov, 1987, Fig. 1), corresponding to the *Cordylodus primitivus* Zone. Deep-water limestones (interval 102 - 107 m) are replaced here by shallow-water ones (interval 107 - 107.5 m), i.e., the drastic short-term transgression forestalled the Lange Ranch regression (Zhemchuzhnikov, 1987). However, it is important to note particularly "the peak" of transgression (interval 103.4 - 104 m – black calcisiltites with micrograded bedding) and "the peak" of regression (interval 107 - 107.5 m – limestones with a fenestral structure indicating an extremely shallow-water environment). Thus, there are already regressive sediments of the Lange Ranch Event within the 104.1 - 107.5 m interval.

An analysis of the conodont associations of the *Cordylodus primitivus* time reveals that at this time, such primitive *Cordylodus* species as *C. primitivus* and *C. andresi* have appeared in transitional and cold-water faunal realms (Dubinina, 1991). Penetration of these species into the warm-water realm took place a little later, since the *Cordylodus proavus* time.

Data on the Batyrbay section demonstrates that the *Cordylodus primitivus* time may be divided into two phases (Dubinina, 1991, Fig. 8). The primitive *Cordylodus* species coexisted in the transitional realm (Batyrbay) with proto- and paraconodonts, *E. oconodontus* (*E.*) *notchpeakensis*, and the more rare (than at the previous *E. alisonae* time) *Teridontus nakamurai* and *E.* (*E.*) *alisonae* during the early phase corresponding to the accumulation of sediments of the 102 - 104 m interval. Analogues of the Batyrbay conodont communities have been recorded in the *Cordylodus primitivus* early phase, in both the warm- and cold-water realms. Conodont communities of the first (warm-water) realm (the epicontinental seas of North America, north China, and Australia) do not differ practically from those of the *E. alisonae* time, whereas the cold-water communities of northwestern Europe have undergone noticeable changes. *C. andresi*, supplemented by proto- and paraconodonts and *E.* (*E.*) *notchpeakensis*, dominates there. *E.* (*E.*) *alisonae*, typical of the transitional realm, has become extinct (Batyrbay) at the end of the first phase. Rare warm-water elements such as *Proconodontus* and *E.* (*Cambrooistodus*) have vanished from syndepositionally redeposited associations of thin turbidites in the Batyrbay section, which can be explained as a natural process of extinction of these species in shallow-water environment. This fact gives rise to suggesting that at the end of the *Cordylodus primitivus* early phase, the warm-water *Proconodontus* and *E.* (*Cambrooistodus*) became extinct in the entire warm-water realm. Perhaps, it was actually connected with the strong drop in temperature, which led to the sharp drop of sea level and, as a result, to the hiatuses in shallow-water sediments of North America, north China, Australia (warm-water realm), and northern Estonia (cold-water realm). Such phenomena undoubtedly corresponded to the Lange Ranch eustatic event, which began, however, not from the *Cordylodus proavus* time, as Miller (1984) suggested, but from the middle of the *Cordylodus primitivus* time.

Communities of the second phase of the *Cordylodus primitivus* time, corresponding to accumulation of sed-

iments of the regressive phase of the Lange Ranch Event at the interval of 104.1 - 107.5 m, are represented by the *Cordylodus* sp., *E.* (*E.*) *notchpeakensis*, and proto- and paraconodonts. It is not possible now to find analogues of the conodont communities of the Batyrbay section not only in the shallow-water basins of the warm- and cold-water realms, but also in the proximal or more or less distal parts of oceanic areas of the transitional realm (western Newfoundland), because of both breaks in sedimentation on the shelves and intense erosional processes on the continental slopes. Perhaps deep-water environments of the cold-water realm (central Kazakhstan, northwestern China) will be more promising in the search for such communities.

Thus, the commencement of the Lange Ranch Event (the middle of the *Cordylodus primitivus* Zone) and faunal changes at the base of the *C. proavus* Zone (at the Cambrian-Ordovician boundary) were not synchronous events (through they were close in time). The first event was succeeded by the second after a very short-time interval corresponding to the second phase of *Cordylodus primitivus* time. Such a discrepancy in time could be neglected on the geological time scale. However, the more precise definition of the Lange Ranch Event commencement itself has principal importance: (1) it confirms the minuteness of research, which has now reached near the Cambrian-Ordovician boundary; (2) it demonstrates that the only section where this minuteness could be established is the Batyrbay section, and, finally, (3) it shows that rearrangement of conodont and trilobite communities near the Cambrian-Ordovician boundary (at the base of the *Cordylodus proavus* Zone) can be regarded as a direct consequence of this event.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 100 - 105.
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Evolution of the biosphere

Trilobite Communities and Facies of the Upper Cambrian Deposits from the Northwest of the Siberian Platform

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Manuscript received March 20, 1992

In 1979, the Cambrian Commission of the Interdepartmental Stratigraphic Committee adopted the stage division of the Upper Cambrian as a biostratigraphic standard for the former USSR territory. The division was elaborated on the basis of southern Khazakhstan sections and included the Ayusokkanian, the Saksian, and the Aksayan stages ("Resheniya...," 1983). These stages have been used in the majority of the Siberian platform Cambrian stratigraphic scales. No serious problems arose in correlation of the sections of the Noril'sk-Kharaulakh facies region with the adopted standard. However, a correlation of the sections of two other facies regions, the Igarka-Kotui and the Turukhansk-Irkutsk, with these stages, is very debatable. The discrepancies between various correlation versions often exceed a stage range (Rozova, 1964, 1968; Datsenko *et al.*, 1968; "Resheniya...," 1983; Zakharov, 1991).

This situation causes considerable confusion in various correlations and questions the relevance of the southern Khazakhstan stages for the subdivision of the Siberian platform Upper Cambrian.

In this respect, it seems very important to use new methods and research directions for a solution to the problem. This paper attempts to reveal correlations between the taxonomic structure and diversity of trilobite communities, on the one hand, and specific lithofacies and facies belts, on the other. Sections of all identified facies regions of the Siberian Upper Cambrian are closely situated in the northwest of the Siberian platform. This is why we concentrated our efforts upon the main sections along the Kureika, Kulyumbe, and Chopko rivers.

We hope the results of this study will enable, to some extent, more accurate understanding of the distribution of trilobite assemblages in the studied area and throughout the section of the Upper Cambrian. We also hope they will contribute to the reconstruction of trilobite successions and to serve as an indirect criterion for estimating interregional correlation correctness for the time interval.

The Late Cambrian biogeographic subdivision of the Siberian platform was proposed by A.V. Rozova on the basis of the study of trilobite assemblages. She has designated the Kharaulakh, Igarka, and Kureika provinces with specific trilobite associations (Rozova, 1979). Boundaries of these provinces are close to those of the earlier defined facies regions (Pisarchik *et al.*, 1967, 1975; "Resheniya...," 1983). In the studied area, A.V. Rozova identified the Norilsk, Igarka, and Kureika trilobite communities, which had inhabited the Noril'sk, Igarka, and Turukhansk facies regions.

Though adopting the facies and biogeographic division proposed, we emphasize the constant and substantial changes in the boundary positions of the facies regions and biogeographic provinces during the entire Late Cambrian. This is why the subdivision proposed for the entire Series is rather tentative.

Before considering lithofacies and associated trilobite assemblages, it is worth noting the general trends of the Late Cambrian sedimentation. It should be mentioned that we assume the base of the *Glyptagnostus reticulatus* Zone as the base of the Upper Cambrian, which coincides with the base of the Chopko Formation in the Noril'sk area. It is also correlated with the base of the Madui Horizon, that is, with the upper parts of the Orakta and Ust'-Pelyadka formations in the Igarka and Turukhansk areas.

By the beginning of the Late Cambrian, a shallow-water carbonate platform was formed in the territory considered. It occupied the Turukhansk and almost the entire Igarka facies region (Fig. 1). The depth of this marine basin increased northwestward, so that in the Noril'sk area there existed comparatively deep-water environments. At this time, the bottom topography was most uneven with the maximum contrast of lithofacies.

As the sedimentation proceeded in the upper part of the slope, the carbonate platform front prograded northward. By the end of the Late Cambrian, the carbonate platform had comprised all of the considered area.

The shelf lagoon was characterized by variable depths (0 - 50 m) and by low or moderate hydrodynamic activity. Here, polycyclic sediments accumulated, represented by mudstones, biolithoclastic wackestones and packstones, oolitic lithoclastic packstones and grainstones, pelloidal packstones, algal limestones, and various dolomites.

In the frontal part of the carbonate platform, with considerably higher hydrodynamics and nearly the same depths, lithoclastic and oolitic grainstones and packstones and algal bioherm limestones accumulated. Polycyclic sedimentation pattern is most likely the main cause of the carbonate platform progradation to the north-northwest. The progradation itself did not favor the formation of reef systems.

The upper part of the slope comprised a marine area with depths varying from the fair weather to the storm wave base and with moderate hydrodynamic activity except for storm events. The main lithofacies of the zone are rocks of the limestone-marl rhythms, crowned by bioturbated limestones, storm breccias, and other tempestites. More rarely, subaqual landslide breccia-conglomerates occur.

In the early Late Cambrian, the northern part of the studied area belonged to a deep-water marine basin, stretching to the north and northeast. Below the storm wave base, pelagic, dark laminated bituminous limestones accumulated, often containing abundant sponge spicules. During regressions, these rocks were replaced by distal and medial calcareous turbidites typical of the lower part of the slope.

This lateral zonation can also be rather distinctly recognized in the vertical direction. As facies analysis is not the goal of this paper, but rather serves as a basis for biofacies analysis, we restrict ourselves to a single example, the Chopko River section (Fig. 2). The lower part of the Upper Cambrian is represented here by the lower Chopko Subformation and corresponds to the *Glyptagnostus reticulatus*, *Erixanium sentum*, and *Glyptagnostotes elegans-Irvingella perfecta* zones. It was formed in the deep-water environment in the lower part of the carbonate platform slope. The middle and upper Chopko subformations, corresponding to the *Tagenarella* Zone and beds with *Tukalandaspis egens*, were formed in an upper part of the carbonate platform slope.

Only the stratigraphically narrow lower part of the Tukalanda Formation corresponds to the frontal part of the carbonate platform, whereas the main part of the formation, comprising the *Ketyna* and *Dolgeuloma* zones, accumulated in the shelf lagoon setting.

Revealing the relations between trilobite occurrences and lithofacies, we establish that the bulk of trilobite remains is associated with specific facies. These are: (1) pelagic microgranular laminated limestones; (2)

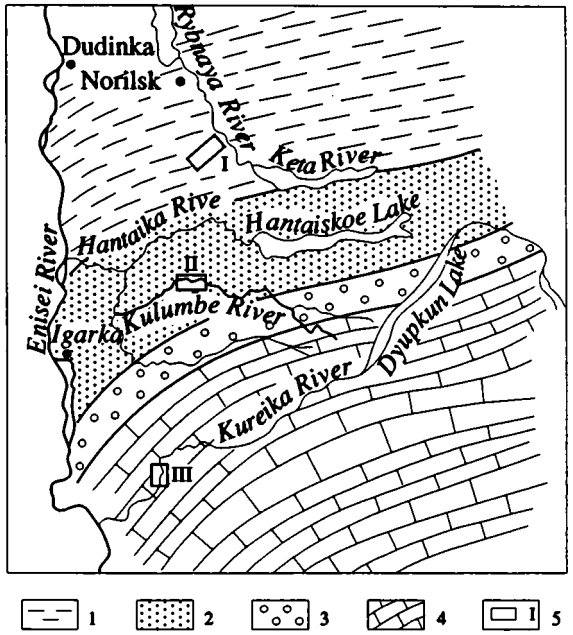

Fig. 1. *Paleofacies subdivision and location of the studied areas.*

1 – facies of lower slope and basin: dark-gray and black mudstones, calcisiltites and marls, calcareous turbidites; 2 – facies of upper slope: gray mudstones, packstones, breccia-conglomerates, coquinas; 3 – facies of carbonate platform margin: light packstones, grainstones, algal boundstones, bioclastic grainstones; 4 – facies of shelf lagoon: light-gray mudstones, wackestones, oolitic grainstones, pelletic packstones, yellow dolomites; 5 – location of sections: I – Chopko River, II – Kulyumbe River, III – Kureika River.

bioclastic and lithoclastic wackestones, packstones, and grainstones, redeposited by gravity flows; (3) storm-generated detrital limestones (bioclastic and, rarely, lithoclastic wackestones and packstones); (4) lithoclastic oolitic grainstones with varying contents of bioclastic material; and (5) algal biohermal limestones.

It is clear from the list presented that the autochthonous burials of trilobites are associated with the first and the last lithofacies only. Trilobite finds, confined to lithofacies 2 - 4, are characterized by a varying degree of transportation and mixture. Therefore, we will later refer to the trilobite assemblages associated with specific facies as taphonomic associations (TA).

Pelagic limestones (Fig. 3a), containing the first trilobites TA , are characterized by black, dark-gray, and green-gray colors, thin parallel bedding, thin flaggy or foliated jointing, and varying degrees of enrichment with organic and clayey matter. There are nearly no signs of material transportation; shell detritus is also absent. The preservation of trilobites in this lithofacies is excellent, and complete specimens are often found. Bedding planes bear trilobite traces. As a rule, the number of specimens in each occurence is insignificant. Taxonomic diversity is also not very high. Commonly, representatives of the Agnostidae superfamily and the

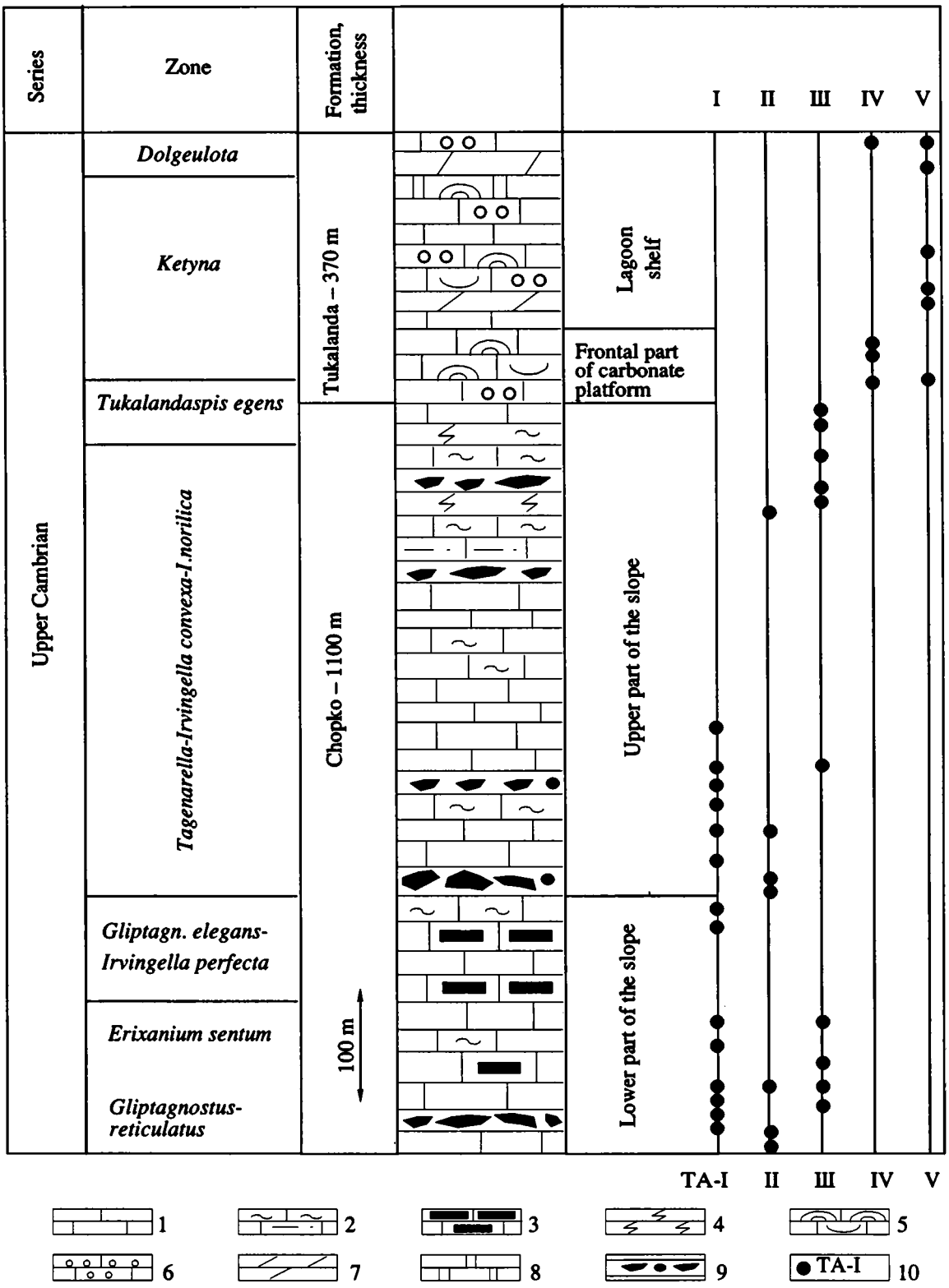

Fig. 2. *The section of the Upper Cambrian deposits along the Chopko River and distribution of trilobite taphonomic associations in the section.*

1 – dark microgranular basinal mudstones; 2 – gray and greenish-gray calcisiltites and marls; 3 – gray mudstones, bioclastic and lithoclastic wackestones and packstones; 4 – gray and light-gray bioturbated limestones; 5 – algal boundstones, brachiopod-trilobite coquinas; 6 – oolitic lithoclastic grainstones; 7 – dolomitic marls; 8 – dolomites; 9 – carbonate breccia-conglomerates; 10 – taphonomic associations of trilobites (I - V) and their position in the section.

Olenidae and Komaspididae families are present. The number of taxa does not exceed 5 or 10 genera or species, but most of them are cosmopolitans. For example, the first TA, stratigraphically confined to the *Irvingella perfecta-Glyptagnostotes elegans* Zone, contains *Olenaspella* aff. *borealica* (Lerm.), *Pseudagnostus securiger* (Lake), and *P. idalis* Öpic, besides the zonal species.

At the level of the *Tagenarella* Zone, the trilobite assemblage of the same TA includes *Parabolinella ?fortunata* Laz. and *Pseudagnostus* aff. *parvus* Sherg.

Bioclastic and lithoclastic wackestones, occurring as blocks and pebbles of gravity flows breccia-conglomerates (Figs. 3b, 4a), are characterized by at least twofold redeposition and certain sorting of fossils. There are essentially no complete specimens or cephalons here, and trilobite remains of the second TA are represented by cranidia and pygidia of various degrees of preservation. The taxonomic composition of trilobites is much more diverse here. For example, from different blocks of the subaqual landslide breccia-conglomerate bed, corresponding to the base of the *Tagenarella* Zone, we have collected the following species: *Irvingella perfecta* N. Tchern., *Onchonotellus trisulcatus* Ivshin, *Convexocephalus granularis* Varl., *Proceratopyge captiosa* Laz., *Entsyna antiqua* Varl., *Pterocephalia* sp. indet., *Tchopkina spinosa* Varl., *T. tchopkinica* Varl. et Rozova, *Eoacidaspis amplicauda* Laz., *Sigmocheilus apheles* Varl., *Pseudagnostus* aff. *mortensis* Sherg., *Heilocephalus* sp., and many others.

A similar taxonomic diversity is observed in other beds of the same type that belong to the *Taganarella*

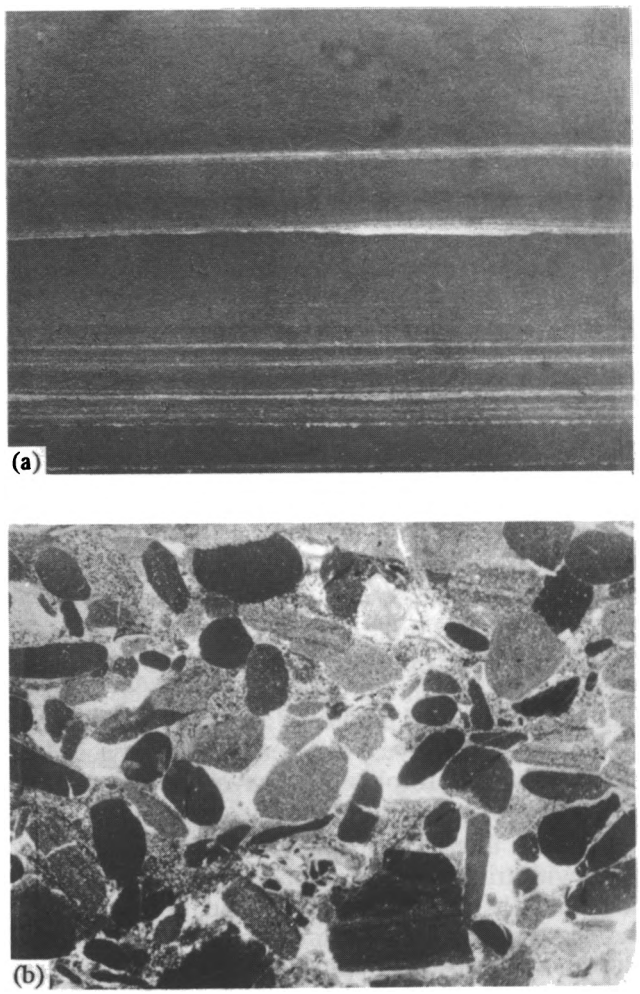

Fig. 3. *(a) Pelagic microgranular limestone with thinnest interlayers of calcisiltite, shell detritus is absent. Enclosing rock of the taphonomic association I; thin section, magnification 8, the Chopko Formation, Chopko River; (b) Lithoclastic grainstone from the block of subaqual landslide breccia-conglomerate. Enclosing rock of a variety of the taphonomic association II, trilobite fragments are very rare; thin section, magnification 8, the Chopko Formation, Chopko River.*

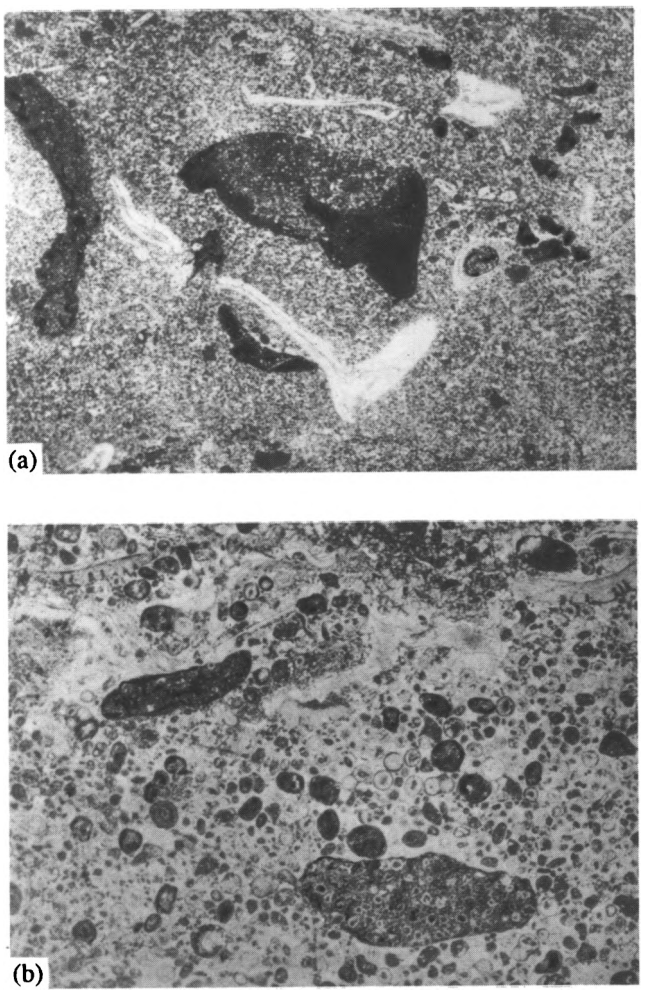

Fig. 4. *(a) Biolithoclastic wackestone with fragments of trilobites and brachiopods, a rock variety associated with the taphonomic association II, thin section, magnification 10, the Chopko Formation, Chopko River; (b) Lithoclastic oolitic grainstone, with abundant fragments of trilobites and brachiopods, enclosing rock of the taphonomic association IV, thin section, magnification 8, the Tukalanda Formation, Chopko River.*

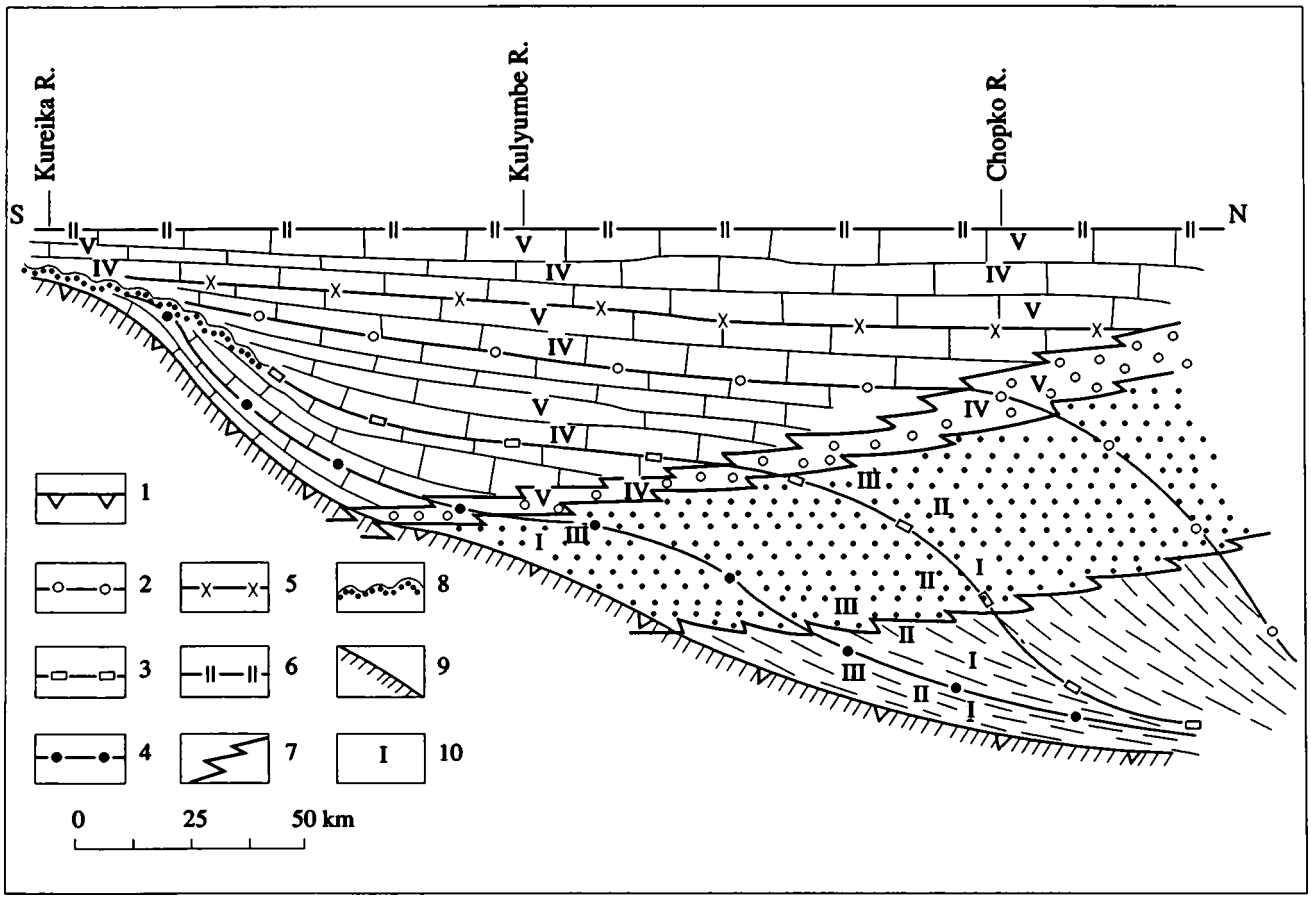

Fig. 5. *Paleofacies profile of the Upper Cambrian deposits in the northwestern Siberian platform and position of trilobite taphonomic associations.*
1 - 6 – Boundaries of biostratigraphic units: 1 – the base of the Madui Horizon (Kulyumbe River) and of the Glyptagnostus reticulatus Zone (Chopko River), 2 – the base of the Entsi Horizon (Kulyumbe River) and of the strata with Erixanium sentum (Chopko River), 3 – the base of Yuraki Horizon (Kulyumbe River) and of the Tagenarella Zone (Chopko River), 4 – the base of the Ketyi Horizon (Kulyumbe and Chopko Rivers), 5 – the base of the Mansi Horizon (Kulyumbe and Chopko Rivers), 6 – the base of the Nyai Horizon (Kureika, Kulyumbe, and Chopko rivers); 7 – facies boundaries; 8 – disconformity; 9 – the Middle Cambrian deposits; 10 – taphonomic associations of trilobites (I - V). Other explanations see Figs. 1 and 2.

and *Glyptagnostus reticulatus* zones within the same lithofacies.

Detrital limestones, mainly reworked by storms, comprise the third TA, which is very similar to the second TA in taxonomic composition and diversity of trilobites. However, it differs in that the preservation of fossils is worse. It is not uncommon for more than 90 percent of the detrital materials to be undeterminable.

Lithoclastic and oolitic grainstones with trilobites of the fourth TA (Fig. 4b) yield indications of the extreme shallow-water conditions and contain variable amounts of detrital material. The number of species is insignificant, no more than 5, and commonly from 1 to 3. The degree of preservation can vary from practically undeterminable remains to well preserved cranidia and pygidia. However, we found no complete specimens in this lithofacies.

Algal bioherms, with which the fifth TA is associated, consist not only of alga remains, but also mainly of lime muds. This indicates a very low hydrodynamic activity. Algal bioherms are characterized by very poor specific composition and satisfactory preservation of trilobites. Occasional complete specimens are present. It is worth noting the very different, most likely original orientations of the trilobites. However, there are practically no fossils in the interbioherm space.

It is clear that we characterized only the basic lithofacies with trilobite associations. However, it is these lithofacies that are most widespread in the sections studied, and the overwhelming majority of trilobite finds used for biostratigraphic zonation are confined to these lithofacies. It should be emphasized that the fabric of these lithofacies and the qualitative and quantitative characteristics of the trilobite assemblages can vary within certain limits. However, the main characteristics considered are surprisingly constant.

In order to elucidate the spatial distribution of the trilobite taphonomic associations, we plotted them on a schematic facies profile, compiled on the basis of the

investigations of the Upper Cambrian deposits in sections along the Kureika, Kulyumbe, and Chopko rivers (Fig. 5). The figure demonstrates that the first and second TA's are confined to the basin and lower slope deposits and are less frequently present in the upper slope deposits. The third TA often occurs in the upper slope and, more rarely, in the lower slope sediments. The fourth and fifth TA's are confined to the carbonate platform proper, both to its frontal and to its shelf-lagoon parts.

The data presented enables us to make the following conclusions:

1. The northwestern part of the Siberian platform was a shallow-water carbonate platform in the Late Cambrian time. In the north-northwest, the platform was surrounded by a relatively deep open-marine basin. In the early Late Cambrian, the frontal part of the platform was situated at the latitude of Igarka. By the end of the Late Cambrian, it prograded to the latitude of Noril'sk, that is, at least 150 - 200 km northward (Figs. 1, 5).

2. The trilobite assemblages of the first, second, and third TA's include 15 - 30 species. They have practically no taxa in common with the fourth and fifth TA's of the same age, comprising no more than 5 - 8 species.

3. The evolutionary changes of trilobite assemblages can be confidently indicated in the case of their occurrence within the same taphonomic association only. The vertical change in trilobite complexes, accompanied by the change of the taphonomic association, may be caused by migration, rather than evolution.

4. The analysis of the change of trilobite assemblages along the section in accordance with sedimentation cycles shows that most biostratigraphic boundaries are confined to the bases of sedimentary cycles, coinciding with the beginnings of transgressions. More rarely, the boundaries coincide with the beginning of regressive sedimentary cycles (Fig. 2). Most likely, there is a natural connection between sedimetological cycles and qualitative renewals of trilobite assemblages.

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Stratigraphy and Geological Correlation, Vol. 1, No. 2, 1993. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 1, No. 2, 1993, pp. 111 - 112.
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The Working Style of V.P. Maslov[1] (to the Centennial of His Birthday)

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Manuscript received Febrary 20, 1992

"Scientific knowledge is an integrated form of spiritual culture, common for all mankind, and, in its base, independent of historic or geographic place and time." (V.I. Vernadsky, 1988, p. 308).

I think that the phenomenon of science could evolve only because outstanding discoveries of scientific thought were concentrated into an integrated whole, owing to an absolutely specific, peculiar working style of many scientists. It is to this working style, which was inherent in many staffers of the Geological Institute of the USSR Academy of Sciences, including Vladimir Petrovich Maslov, that I should like to draw your attention. I hope that my notes will help to comprehend this particular working style, which was so characteristic of Professor Maslov.

In my life, I had fortunate opportunities to meet interesting people. But among them, Vladimir Petrovich, who was an astonishing person, takes an absolutely special place. I first became acquainted with the name of Vladimir Petrovich Maslov in 1953. In that year, I became a postgraduate of Leningrad University and began my studies of Middle Paleozoic foraminifers. I took great pains to acquire the first edition of Professor Maslov's *Atlas of Rock-Forming Organisms*. The book indelibly impressed me with its tremendous scope of information and the simplicity of its presentation. Therefore, six years later, when I wrote my first report on the results of my studies of Devonian charophytes from the Tien Shan, it was to Vladimir Petrovich that I felt a desperate desire to send the report so that he could revise it. The point was that in those years, umbellas were attributed to foraminifers, but I personally came to the conclusion that they were peculiar Charophyta! And, as I thought, it was only Vladimir Petrovich who could decide whether my conclusion was correct and that I was right in contradicting the generally accepted view.

A copy of my report was sent to him. After a month, I received it back with a very benevolent and good review that supported all my conclusions. There was also a letter attached to it, along with a stack of small sheets of paper which contained comments, mainly on the terminology. The letter favorably assessed my work: V.P. Maslov wrote that after a few additional touches, my work might become a Candidate dissertation. By that time I had already received that title, so the question of the dissertation was no longer relevant. But we began writing to each other and then became personally acquainted after our meeting at the Geological Institute (GIN). In those years, the GIN was a prominent institution that was universally recognized in a special way because working there simultaneously were such outstanding persons as V.P. Maslov, D.M. Rauzer- Chernousova, V.V. Menner, and N.S. Shatskii.

A day came when I had my first chance to examine my thin sections together with Vladimir Petrovich. I remembered that day by the following discussion of ours: "Don't you think, Budimir Vladimirovich, that if the thin section is oriented in just this way, the cross-section will look more natural?" I turned the thin section by 180 degrees. And, indeed, the most amazing things become opened to my sight, to the accompaniment of Vladimir Petrovich's explanations. "At the top you can see a small apical lid, and at the bottom, a basal plate, a basal pore, and basal thickened growths." There are two moments in this fugitive scene between the Master and the young, absolutely green scholar, to which I should like to draw your attention.

First, the elder colleague demonstrated the highest possible delicacy when addressing his younger partner with the words "Don't you think...." Such a style of communication definitely penetrates to one's soul forever. Later, I observed the same style of communication when talking with V.V. Menner. Their style drastically contrasted with the manner of criticism existing in those years. It was owing to V.P. Maslov, D.M. Rauzer-Chernousova, and V.V. Menner that such a style of communication was perceived by me as that typical of the GIN in those years.

[1]The text is based on the author's oral presentation made on May 21, 1991 to the session of the Moscow Society of Naturalists, Geological Institute of the Russian Academy of Sciences and the Council "Problems of Paleobiology and Evolution of the Organic World," dedicated to the centennial of V.P. Maslov's birthday.

Second, V.P. Maslov had a unique skill in noticing the most minute structural details, which he could splendidly use for his reconstructions. I simply admired this skill of his. It goes without saying that to be so skillful, one must be a tremendous connoisseur of world experience but, besides that, must know a great many of other things. And, most importantly, he must know what to look for in the particular as based on the general.

I was always very fond of working with two of V.P. Maslov's monographs, namely *The Fossil Calcareous Algae* and *The Red Algae*. Their descriptions differed greatly from the canonical order of describing various species. The figures accompanying the text helped one to comprehend its meaning, organically merging with the text. I was fond of getting to the core of the material written by the author, speculating on why he chose that particular way of writing and not any other. I was enchanted by the wonderful integration of the text and the figures that were all so splendidly made by V.P. Maslov himself. The feeling was that a living thought, breathing and sparkling, was coming to you from the pages of his books. And that thought had analysis and synthesis coalesced in it.

Once, on learning by chance about my planning to go to L'vov, V.P. Maslov drew my attention to the plates that paved to streets and staircases in L'vov. "This is one of the richest collections of red algae, which has opened a lot for me and has taught me a lot, too." I was slightly shocked – to draw knowledge for scientific research, for monographs, from streets and pavements! Then, after I carefully scrutinized those stone plates, I realized how justified Vladimir Petrovich's words were. Indeed, the plates presented the most abundant and the most diversified material polished not in the laboratory but by many hundred thousands of feet. It really provided the richest food for thought. Not a single collection made up by any researcher's hands could be compared to that one.

The next case I am going to tell you about is associated with my studies of foraminifera systematics. There were some forms whose affinity I could not identify for sure. It was difficult to say if they belong to algae or to foraminifers. No matter how thoroughly I examined their morphology, it could not give me any unambiguous answer. Then I turned for help to V.P. Maslov. "What am I to do?" I asked. He replied as follows. "Look at the problem from a wider angle. If the morphology fails to give you an unambiguous answer, pay attention to the peculiarities of occurrence of the forms that interest you. Try to see whom they are together with most frequently, with the forms beforehand considered as algae or as foraminifers. Find your solution through their environmental distribution and consider your problem in the light of the integrated picture of the past." His advice actually helped me. And he was so helpful not only in that case alone. The ability to see a problem from different angles and to embrace different scales in one study was very characteristic of V.P. Maslov's working style. This aspect of his work is still to be properly appreciated by us. Vladimir Petrovich used to convey to others his own approaches to research, his own experience and his own understanding of things. But he always did it without any obtrusion, by making polite comments and expressing his delicate wishes. Owing to that, everything he said was sure to be driven to your heart and mind for good.

There is another specific feature of V.P. Maslov's working style, which I noticed somewhat later. It was at the time when the volume on algae, which was part of *Treatise on Paleontology*, and the second edition of his *Atlas of Rock-Forming Organisms* were published. In those books, he wrote not only essays on such groups of fossils that he undoubtedly knew so well, but also sections of the books that dealt with the least studied and "most difficult" groups of fossils. All of this permitted judgement about the greatest scope of Vladimir Petrovich's intellect and emphasized the encyclopedic depths of his knowledge. And his own explanation was very simple: "There was just no expert to enlighten that problem, so I had to write about it myself." Now I see that such an approach expressed Professor Maslov's particular aspiration to present a whole, integrated picture of the natural phenomenon he was describing, be that fossil algae or rock-forming organisms.

I had a fortunate opportunity to take part in writing two collective monogfaphs. The work was guided by V.P. Maslov, who was actually the soul and heart of it. These were the *Fossil Charophyta* and the second edition of the *Atlas of Rock-Forming Organisms*. Our joint work opened some new features of this wonderful person. First, he proved to be extremely thoroughly, capitally trained and prepared for collective work. Before our collective book, *Fossil Charophyta of the USSR*, was published, he personally issued a special paper entitled *Introduction to Studies of Fossil Charophyta*. Thus, he seemed to have set up a methodological basis for our joint efforts. Second, Vladimir Petrovich demonstrated his tremendous respect for our future readers. When working with our texts, he did his level best to make them completely clear and convenient for the readers. He paid special attention to compiling tables and writing legends to pictures and photographs. The general trend of science in those years was to emphasize the importance of new discoveries, or to put it in other words, of "riding into the unknown." That was regarded as the main goal of science. In his approach to editing, Vladimir Petrovich displayed somewhat different accentuation. His ideas, at least for me, became clear later. To obtain new knowledge, V.P. Maslov thought, is not the final goal, the final result of investigations, but only an intermediate means to initiate a new action, a new activity. I first felt what he meant when communicating with him, but my real comprehension of that meaning, which I could put to words, came to me much later.

I had always had a feeling that V.P. Maslov was among those who carefully cherished traditions; that in having contacts with him, I was contacting my own precursors. But today I can only say that in his style of thinking, in his approaches, in his great skill in so organically combining thorough analysis and splendid synthesis, V.P. Maslov was in many ways far ahead of his time. It was not always that those around him could properly understand him. Nevertheless, he never uttered a single word of criticism and only smiled sweetly at any objections. His smile seemed to say: "A time will come and you will understand the meaning of the things I am doing now." I had never heard from V.P. Maslov about the interdiscipline synthesis or the activistic approach. But recalling the days of our brief and infrequent meetings, I unwillingly come to the idea that my communication with him had prepared my mind for comprehending such approaches. All our meetings left in my soul the feeling of something great and very bright. Maybe it was due to the fact that throughout all of his life and his activities, Vladimir Petrovich used to strive to the three greatest values existing in the world, namely, truth, kindness, and beauty. May our thanks rest with him.

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